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PREFERRED ORIENTATION OF PEBBLES IN SEDIMENTARY DEPOSITS

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ABSTRACT

Pebbles from glacial till, from an outwash terrace, and from a beach show characteristic preferred orientations. Methods used in collecting and measuring the pebbles are described, and the importance of two distinct concepts in orientation analysis is stressed: (1) the dip direction and dip of the longest axis and (2) the orientation of the maximum projection plane of the pebble. Methods of plotting the data are discussed, including conventional petrofabric diagrams, polar co-ordinate charts, and histograms based on azimuthal and dip frequencies. The principles of statistical analysis of the data by radius-vector summation and by conventional moment analysis are developed. The latter affords numerical data suitable for comparative studies of size, shape, and other attributes with preferred orientation, and permits the development of isopleth maps showing areal variations of preferred orientation.

INTRODUCTION

The successful results of petrofabric analysis of igneous and metamorphic rocks point the way to equally significant results from sedimentary data. The techniques applied to igneous and metamorphic rocks are well established, and methods of graphic presentation by means of petrofabric diagrams are standardized.[1] Among the scattered papers on the orientation analysis of sedimentary particles, especially clastics, techniques and graphic methods have varied widely. Richter,[2] for example, used a field technique for ob-

[1] E. B. Knopf and E. Ingerson, "Structural Petrology," *Geol. Soc. Amer. Mem. 6* (1938).

[2] K. Richter, "Die Bewegungsrichtung des Inlandeises, rekonstruiert aus den Kritzen und Längsachsen der Geschiebe," *Zeits. f. Geschiebeforschung*, Vol. VIII, No. 1 (1932).

serving the orientation of pebbles in glacial till and presented his data as histograms; more recently[3] he published several petrofabric diagrams. Wadell[4] developed a precise sampling and laboratory technique for pebbles and used polar co-ordinate paper to present his original data. The data were summarized as histograms for interpretation. Holmes[5] also used a field technique for studying the orientation of till pebbles and presented his data in modified star diagrams. A comprehensive paper on particle orientation was recently published by Cailleux,[5a] in which the degree of orientation of the long axes of pebbles was designated as "good, fair, or poor" with respect to a mean direction. Cailleux includes a fairly comprehensive bibliography in his paper.

Orientation analysis of clastic sediments may require laboratory methods different from those of conventional petrofabric analysis. Crystal axes or faces are of less significance than the orientation of long axes or other reference lines or planes in the particles, although the crystallographic properties may be related to the relative dimensions of the grains. Moreover, conventional petrofabric diagrams cannot readily be used for direct comparison with other sedimentary characteristics because the data are not summarized numerically. On the other hand, some statistical parameters fail to disclose details of sediments which may best be brought out by standard diagrams.

The writer's interest in the subject was stimulated by Richter's work on glacial till and Wadell's development of an exact laboratory technique. In addition, and from perhaps a broader viewpoint, it is desirable to have quantitative methods for all the characteristics of sediments which can be measured. Among the attributes of sediments, those of size, mineral composition, and shape are supported

[3] "Ergebnisse und Aussichten der Gefügeforschung im pommerschen Diluvium," *Geol. Rundschau*, Vol. XXVII (1936), pp. 197–206.

[4] H. Wadell, "Volume, Shape, and Shape Position of Rock Fragments in Openwork Gravel," *Geografiska Annaler, 1936*, pp. 74–92.

[5] C. D. Holmes, "Till Fabric," *Bull. Geol. Soc. Amer.*, Vol. XLIX (1938), pp. 1886–87 (abst.). Additional information from personal conversation.

[5a] A. Cailleux, "La Disposition individuelle des galets dans les formations detritiques," *Rev. de geog. phys. et de geol. dynam.*, Vol. XI (1938), pp. 171–96.

by quantitative methods of study. Orientation analysis and surface-texture analysis have lagged behind the others. It is the purpose of this paper to discuss possible methods of graphic and statistical analysis which apply to orientation data, as well as specific sampling and laboratory methods suitable to certain size ranges. The techniques will be confined to pebbles because of their ease of manipulation; graphic and statistical methods, while illustrated with pebble data, will of course apply to similar data from any size range.

COLLECTION OF ORIENTED SAMPLES

Wadell[6] used a Brunton pocket transit as a prismatic compass to collect pebbles from an esker and a delta. Vertical and horizontal lines were drawn on the pebble with red and black lacquer, respectively, while the pebble remained in its original position in the outcrop; and the pebbles were taken to the laboratory for further study. The writer simplified the field technique by using a rectangular wooden frame fitted with brass rods and a spirit level. The essential features of the device are shown in Figure 1. The frame is an ordinary 5 × 6-inch photo-printing frame with glass and back removed. The sides were center tapped, and thin brass rods were inserted. The rods were grooved at the point of intersection to lock them into position with each other. A small spirit level was fastened to the lower right side of the frame and calibrated so that it is level when the horizontal brass rod is level. In collecting the pebbles, the frame is held upright, level, and parallel to the face of the exposure, with the crossbars in front of the pebble. A pencil line is drawn on the pebble in the upper right quadrant of the frame, to mark an L on the pebble. This L is a right angle when the pebble is viewed with the eye at the intersection of the rods in the frame. If the surface of the pebble is cleaned with the finger before marking, the pencil line remains clear during subsequent transportation to the laboratory.

In detail the procedure involves choosing an outcrop area about 2 feet square (depending upon the abundance and size of pebbles in the deposit) and removing all loose pebbles from the face. The trend of the cleaned face is carefully determined with a compass and re-

[6] *Op. cit.*

corded as an azimuth measured clockwise from the north point.[7] The dip and strike of the sedimentary bed are also recorded. The pebbles are systematically collected, working over the entire face; if the material is glacial till, for example, the exposed but embedded pebbles are first marked and collected, and the next deeper layer of pebbles is exposed by scraping the fine material away with a broad

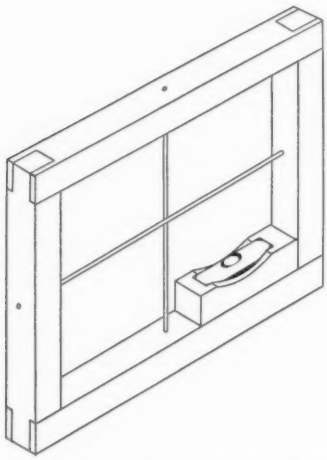

FIG. 1.—Frame for collection of oriented pebbles

knife blade. The individual pebbles are collected by holding the orienting frame as described earlier and marking the L's on them. During the collection of the pebbles it is important that the frame be held always parallel to the collecting face, inasmuch as that direction is the fundamental reference line. The L's are marked in such a manner that the pencil lines extend around edges and curves, so that the figure becomes a right angle only when viewed from one position.

[7] The subsequent laboratory procedure is simplified if the trend of the collecting face is expressed as the azimuth on the observer's left as he faces the outcrop. For example, the collecting face trends N. 30° W., and the observer faces northeast. The azimuth on his left is 330° measured clockwise from north; on his right the azimuth is 330° − 180° = 150°.

LABORATORY STUDY OF THE PEBBLES

In the laboratory the pebbles are brushed free of sand and clay with a small finger brush. The penciled lines are carefully traced with India ink or quick-drying lacquer so that the marks will not be destroyed by subsequent handling.[8] Either or both of two attributes of the pebbles may be studied in orientation analysis. The commonest choice is the longest axis through the pebble (Wadell, Richter, and others); but there are really two independent variables in orientation work, and a complete study should perhaps include both.

1. OPERATIONAL DEFINITIONS OF PEBBLE AXES

Orientation analysis demands precise definitions of pebble axes and planes so that the data may be objective and reproducible. The longest axis of the pebble, most commonly used in orientation studies, is readily found with a sliding block of the type illustrated in Krumbein and Pettijohn's *Manual*.[9] The pebble is rotated in the device until the longest axis is found, and the points of emergence of this axis are marked on the pebble with pencil or lacquer.

The longest axis may be used alone in orientation studies, but the complete orientation of the pebble can be uniquely stated by this axis and a plane. The writer uses the longest axis and the maximum projection plane of the pebble, the latter expressed by its pole, to express the orientation of the particles. The axis and plane are determined operationally by means of a photoelectric-cell apparatus, described in detail elsewhere.[10] The device permits the location and measurement of three mutually perpendicular axes. Two of these, the longest and intermediate (designated as the operational *a*- and *b*-axes, respectively), define the maximum projection area (plane) of the pebble; the intermediate and short axes (operational *b*- and *c*-axes) define the minimum projection area of the pebble. The *c*-axis

[8] If the original lines are drawn with a well-sharpened pencil, of grade 2H to 4H, they remain clear remarkably well. Until the information is all assembled, however, it is well to guard against loss of critical data by lacquering the lines, which may require about an hour's time.

[9] W. C. Krumbein and F. J. Pettijohn, *Manual of Sedimentary Petrography* (New York: D. Appleton–Century Co., 1938), p. 145.

[10] Krumbein, "Application of Photoelectric Cell to the Measurement of Pebble Axes for Orientation Analysis," *Jour. Sed. Pet.*, in press.

is normal to the maximum projection plane, and the *a*-axis is normal to the minimum projection plane.

The photoelectric device consists of a metal support in which the pebble is held with its long axis vertical. The pebble intercepts a parallel beam of light which falls on a photoelectric cell. A milliammeter attached to the cell through an amplifying unit records the amount of light which passes the edges of the pebble. As used, the ammeter reading is a function of the area of the shadow cast by the pebble. The pebble is rotated about a vertical axis until the reading is a minimum, whereupon the maximum projection area of the pebble lies normal to the beam of light. The *b*-axis is defined as the maximum horizontal width of the pebble shadow in this position. The pebble is then rotated 90° to obtain the length and points of emergence of the short *c*-axis, which is defined as the maximum horizontal width of the shadow in the new position.

In the routine adopted by the writer the pebbles are first placed in the sliding block to locate the longest (*a*) axis. The ends of this axis are marked on the pebble. The pebble is then mounted in the photoelectric apparatus, and the points of emergence of the shortest (*c*) axis are likewise marked on the pebble. As a result, two important reference features of the pebble are known. The longest axis lies within or parallel to the maximum projection plane, and the shortest axis is normal to the maximum projection plane. The short *c*-axis is used as a pole of the maximum projection plane to express its orientation in space, and the long axis indicates specifically how the pebble lies with respect to this plane.

The independence of the long axis and the maximum projection plane, and the need for both to orient the pebble uniquely, may be seen by considering any triaxial pebble in a sedimentary deposit. For a given orientation of the long axis the maximum plane may be rotated through any angle about that axis. For example, with a low dip of the long axis the maximum plane may be essentially horizontal ("flat-lying" pebble), it may be essentially vertical (pebble "on edge"), or it may lie somewhere between. Likewise for a given orientation of the plane, the long axis may have various orientations. If the maximum projection plane is inclined, the long axis may be horizontal, or it may dip at any angle up to the maximum angle of

inclination of the plane. Thus, neither the long axis alone nor the maximum plane alone is sufficient to describe completely the position of the pebble in space, except in certain restricted shapes, as some disks and cylinders.

II. MEASUREMENT OF PEBBLE ORIENTATION

Wadell[11] developed a rigorous method for measuring the orientation of the long axis with a two-circle goniometer. His technique is followed here exactly but is supplemented with a recording form which yields the azimuth of the reference lines in the pebble without tedious computations. Wadell's method is adaptable to the measurement of the short *c*-axis also, and the two can be read in a single mounting of the pebble.

The pebble is mounted on the crystal holder of the goniometer with putty or plasticine. By keeping the eye on a level with the pebble, the latter may be manipulated until the penciled or lacquered lines appear as a right angle oriented like an L. The final position is checked by holding the orienting frame of Figure 1 before the pebble, parallel to the vertical circle of the goniometer. In this manner the pebble is mounted in the same relative position as it occupied in the outcrop, with the vertical circle now representing the trend of the outcrop face. The steps involved in reading the orientation may be standardized as follows:

1. Mount the pebble as described.
2. Read the value of the marker on the horizontal goniometer scale.
3. Rotate the horizontal stage *counterclockwise* until the long axis is parallel to the vertical circle of the goniometer, in such manner that the long axis dips *from right to left*. This is checked by looking downward over the vertical circle and observing the positions of the lacquer dots with respect to the vertical circle.
4. Read the value of the marker on the horizontal goniometer scale.
5. Hold a pencil in front of the pebble, inclined so that it coincides with the lacquer dots. Read the inclination of the pencil with the rod and shoe on the vertical goniometer scale. This records the dip of the long axis.
6. Rotate the horizontal circle to the original reference mark of operation (2), and repeat the observations for the short *c*-axis of the pebble.

The data for each reference axis are recorded in Table 1.

In the examples given, the strike of the outcrop face is N. 30° W., as measured on the observer's left as he faces the outcrop. This is

[11] *Op. cit.*

equivalent to an azimuth of 330°, measured clockwise from the north point. The first pebble was placed in the goniometer, and the horizontal scale reading was 281°. Counterclockwise rotation to parallel position yielded a new scale reading of 293°. The difference, 12°, is added to the azimuth of the outcrop face (330°) to obtain 342°. Inasmuch as this value is less than 360°, it is transferred directly to column 6. The angle of dip of the axis is entered in the last column. In the second example the pebble is rotated 167° to parallelism; and this value, added to 330°, yields 497°, which is larger than 360°. Hence, 360° is subtracted from the value in column 5 to obtain the corrected azimuth, 137°. The dip of the axis here was 27°. To avoid confusion, separate tables should be prepared for the long and short axes.

TABLE 1

Pebble Number (1)	First-Stage Reading (2)	Second-Stage Reading (3)	Rotation in Degrees (4)	Rotation Added to Azimuth of Outcrop (5)	Corrected Azimuth (6)	Dip of Axis (7)
1	281	293	12	342	342°	18°
2	153	320	167	497	137°	27°

The foregoing operations involve essentially the following analysis. After the pebble is mounted, the vertical circle represents the trend of the outcrop face. The problem is to find the orientation of the axis in space independent of the accidental trend of the collecting face. The rotation of column 4 indicates how large the departure of the dip direction is from the trend of the face; the addition of the azimuth of the face then converts this to an orientation with respect to the north point. The particular choice of operations is arbitrary; these were designed to afford a method almost automatic in nature.

It is possible to reduce the number of readings by making a brass mount for the pebble. A circular platform about an inch in diameter, with a brass pin in the center, is inserted in the crystal support of the goniometer. This mount is free to rotate without disturbing the horizontal scale reading. The pebble is mounted on the

brass base, with the horizontal marker left at 0°. When the pebble is rotated into parallel position, the horizontal scale reading is that of column 4 directly.

GRAPHIC PRESENTATION OF THE DATA

The compiled data of Table 1 may be presented in various manners. Wadell used polar co-ordinate paper; by and large, this is probably the most satisfactory manner, because it supplies the original unmodified data for later reference. The tendency among structural petrologists is to publish petrofabric diagrams directly, without including the original distribution of axes. Petrofabric diagrams are essentially interpretative devices; it would be advantageous if they were supplemented with polar co-ordinate graphs of the original data, so that later workers could use the original observations for other purposes if they wished.

Petrofabric presentation involves the use of an equiareal net (the so-called "Schmidt net") to plot the points in proper relation to each other for contouring. This suggests the use of equiareal polar co-ordinate paper for the original data, so that the points themselves are in the same relative positions that they occupy in the petrofabric diagram. Hence, if one prefers to publish a single diagram, such a chart would afford the original data for those who wished it, and, at the same time, would supply the points properly shown for sketching contour lines without the necessity of redrawing the figure.

Equiareal polar co-ordinate paper differs from ordinary polar co-ordinate paper only in the spacing of the concentric circles. Figure 2 shows a portion of ordinary paper on the left and a portion of equiareal paper on the right. The azimuthal readings along the circumference are not modified; the only difference is in the spacing of the angles of dip. The three similar points plotted on the charts indicate the "foreshortening effect" of the equiareal paper. This paper was prepared from the Leitz 10-cm. equiareal stereographic net, and all subsequent figures in this paper are plotted on it.

Plotting the data involves two co-ordinates: the azimuth of dip direction (Table 1, col. 6) and the angle of dip (Table 1, col. 7). Azimuth is measured clockwise around the circumference, and the angle of dip increases from the circumference toward the center.

The plotted points therefore show the direction and inclination of the axes. A point at azimuth 350°, dip 10° (Fig. 2) means that the axis is plunging downward at 10° in a N. 10° W. direction. This convention of plotting conforms to that adopted by structural petrologists.

This discussion of graphic presentation is not to be construed as a criticism of petrofabric diagrams as interpretative devices. The wide and successful use of such diagrams in structural petrology suggests a similar usefulness with sedimentary data. However, the

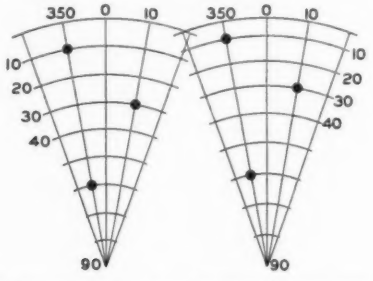

FIG. 2.—Relation between ordinary polar co-ordinate paper (*left*) and equiareal polar co-ordinate paper (*right*).

polar co-ordinate chart supplies numerical data which cannot be derived directly from the petrofabric diagram. These numerical data open new fields of description and comparison of the data, which do much to supplement the information obtained from the petrofabric diagrams.

As an example of the detail with which orientation data may be studied, a sample of one hundred glacial till pebbles was analyzed in terms of size, shape, and preferred orientation of the long *a*-axes and short *c*-axes. Figure 3 shows the original orientation data; the diagram at the left is for the long *a*-axes, and that at the right shows the short *c*-axes of the same sample. The *c*-axes represent the poles of the maximum projection planes, and the two diagrams present the factual data of the till fabric.

The pebbles were collected from a road cut through a late Wisconsin drumlin near the town of Random Lake in eastern Wis-

consin. The drumlin trends S. 74° W. and is one of a field of drumlins whose average trend is S. 82° W. The ice presumably traveled from east to west from the Lake Michigan basin on the east. The distribution of points in the left diagram of Figure 3 shows a concentration in an east-west direction, indicating statistically that the long *a*-axes tend to parallel the direction of ice movement, as Richter established some years ago. The distribution of *c*-axes in the right diagram is approximately north-south, or normal to the direction

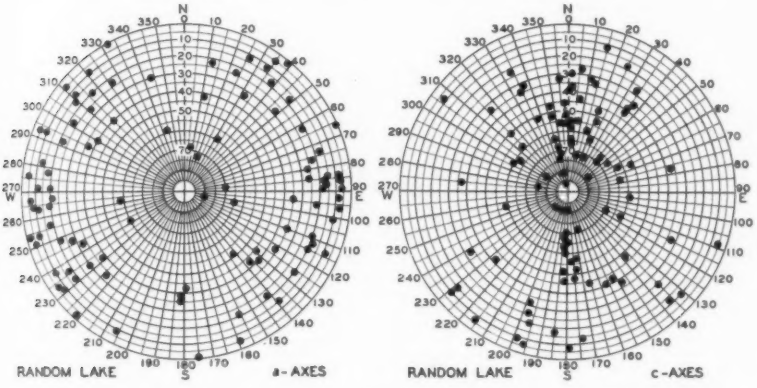

FIG. 3.—*Left*, distribution of long *a*-axes of late Wisconsin till pebbles. *Right*, distribution of short *c*-axes of same sample.

of ice movement. The *c*-axes are the poles of the maximum projection planes of the pebbles, and the interpretation is that the pebbles tend to present their minimum cross-sectional areas opposed to the direction of ice flow, so that statistically the maximum cross-sectional areas lie parallel to the direction of movement.

The clustering of points in the *c*-axis diagram is noticeably better than for the *a*-axes. Although the *c*-axes are perpendicular to the *a*-axes by definition, it does not follow that the azimuths of the two axes are perpendicular to each other. This can readily be illustrated, as shown in Figure 4. Consider a pebble whose *c*-axis is oriented with an azimuth of 330° and a dip of 40° (*P* in Fig. 4). The relation between the *a*- and *c*-axes is that the *a*-axis must lie in a plane normal to *P*. This plane is represented by the curve *CAB*.

If the long axis is horizontal, it will emerge at point B, and the azimuths of the two axes will be at 90°. If the a-axis dips, say 25°, it will emerge at A, and the azimuths of the two axes will no longer be perpendicular. (A has azimuth 80°, 110° from P.) Hence, the two diagrams of Figure 3 show independent concepts, except for the condition that one axis must always lie somewhere within a plane at 90° from the other axis.

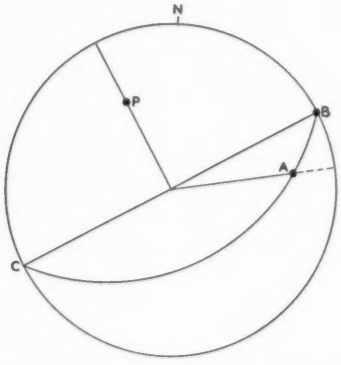

FIG. 4.—Equiareal projection of pole P, its normal plane, and position of axes within plane.

Petrofabric diagrams of the till pebbles disclose the same data somewhat more graphically (Fig. 5). These diagrams were prepared according to directions in Knopf and Ingerson's *Memoir*[12] using a 1 per cent circle. The diagrams are oriented with the north point at the top, and a line through the points AA indicates the trend of the drumlin axis. This line also indicates the approximate position of the a-fabric axis (the axis of movement). The symmetry of the diagrams is monoclinic, as may be anticipated in depositional fabrics. The gentle dips of the a-axes, and their orientation in the direction of ice movement, are in striking contrast to the steeper dips of the c-axes, which tend to be normal to the direction of ice movement.

Neither the polar co-ordinate diagram nor the petrofabric diagram can be used directly to express the average direction of

[12] *Op. cit.*, pp. 245–49.

orientation of the axes numerically. The petrofabric diagram suggests a mean direction slightly north of east for the *a*-axes and an approximately north-south direction for the *c*-axes. To what extent the degree of orientation is better in one diagram than in the other cannot be read directly, although the distribution and size of the black areas suggest less spread for the *c*-axes.

For numerical data, a rigorous location of the mean direction may be had by using standard radius-vector summation methods, pro-

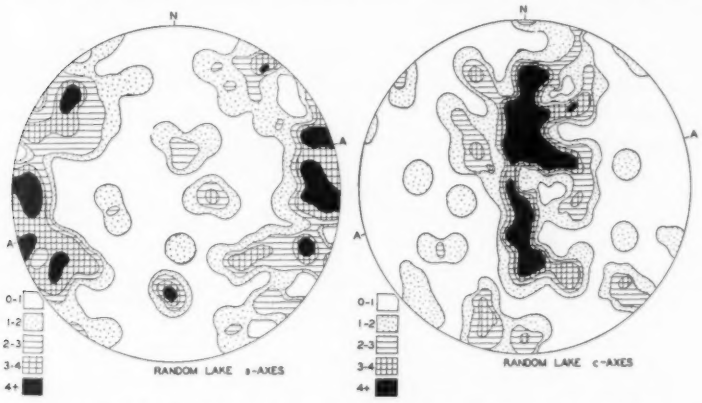

FIG. 5.—Petrofabric diagrams of *a*-axes of till pebbles (*left*), and *c*-axes (*right*)

vided there is any preferred orientation shown by the data. The rather tedious computations of radius-vector summation may be avoided by using conventional moment methods of analysis, with an additional advantage that not only the mean direction but also the degree of spread and the symmetry of the distribution may be expressed numerically.

STATISTICAL ANALYSIS OF ORIENTATION DATA

Petrofabric diagrams combine two independent variables into a single composite picture. These variables are the direction of dip and the angle of dip of the axes. The complete picture of the fabric obviously includes both, because in a sense the data are vectorial. A vector is a directed line segment; it has both direction and magnitude. An oriented pebble axis may be expressed as a vector by draw-

ing its length proportional to its angle of dip. Essentially this approach was made by Reiche in his study of the Coconino sandstone,[13] using the poles of the cross-laminae as the vectors. In some studies it may be advantageous to separate the variables and to discuss the mean azimuth of all axes independently of their dip, and vice versa. To do this, the azimuths are drawn from the original polar diagram as a frequency distribution independent of dip. Similarly, the dip distribution may be drawn from the diagrams independent of the azimuths.

I. RADIUS-VECTOR SUMMATION METHOD

As a first approach to the statistical analysis of orientation data in terms of azimuth alone, it is necessary to group the points into classes. Any choice is permissible; but for convenience the class

TABLE 2

FREQUENCY TABLE OF TILL-PEBBLE *a*-AXES

Class Mid-point	Frequency	Class Mid-point	Frequency	Combined Frequency
0°	1	180°	4	5
20	5	200	1	6
40	8	220	2	10
60	2	240	11	13
80	12	260	10	22
100	8	280	6	14
120	6	300	7	13
140	5	320	7	12
160	2	340	3	5

intervals may be chosen as 20° each, starting with azimuth 0° as the mid-point of the first class. This defines the classes as 350°–10° (mid-point 0°), 10°–30° (mid-point 20°), etc. By counting the number of points in each class of the till-pebble *a*-axes of Figure 3, the data of Table 2 were obtained. The classes are expressed by their mid-points. The last row of the table shows the sum of the points in corresponding classes opposite each other on the circle. A histogram of the eighteen classes is shown in Figure 6. There are at least two principal modes which are strikingly brought out in the inset

[13] P. Reiche, "An Analysis of Cross-Lamination: The Coconino Sandstone," *Jour. Geol.*, Vol. XLVI (1938), pp. 905–32.

figure, which gives the same data combined into 40° classes. The modes of the inset occur at approximately 90° and 270°; inspection of Figure 3 indicates that this agrees with the trend shown by the points. In terms of statistical analysis, however, this bimodal arrangement of the data interferes with the location of the mean preferred orientation. That is, in terms of azimuth alone, classes on opposite sides of the circle are equivalent; and unless they are combined, the one cancels the other to the degree that the number of points in each opposite class is equal. Thus, it is not adequate in

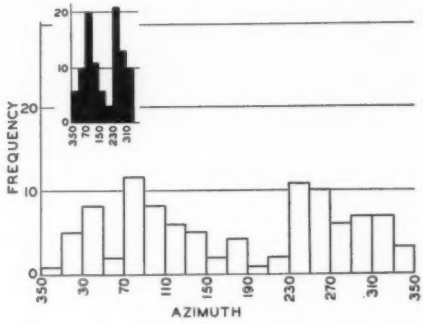

FIG. 6.—Azimuthal distribution of till pebble *a*-axes over 360° range. Inset, same data in 40° classes.

the general case to compute the mean by such a formula as $(1/N)(\Sigma f \cdot \theta)$, where N is the total frequency, f is the frequency in each class, and θ is the class mid-point. The data of Figure 6 yield a mean of approximately 180° by this method, which is obviously about 90° in error, because it falls in the lowest part of the bimodal curves of Figure 6.

It will be seen later that the method which fails in this case may apply in special circumstances, but for the general case a rigorous method of finding the mean must be adapted to periodic data. Such a method is the radius-vector summation method,[14] which consists essentially of combining the data into azimuthal classes such that the class $(\theta + 180°)$ is added to the class θ. This yields a distribution

[14] Suggested to the writer by Dr. T. L. Page, of the Department of Astronomy of the University of Chicago.

extending over 180°; and to maintain the periodic nature of the distribution, the angle associated with each class mid-point is doubled. A radius vector is associated with each doubled class mid-point, having a length proportional to the frequency in the class. This yields a set of radiating lines, as shown in Figure 7, *A*, which roughly resembles a lily pad. Each radius vector is then resolved into two components in terms of the sine and cosine of the angle,

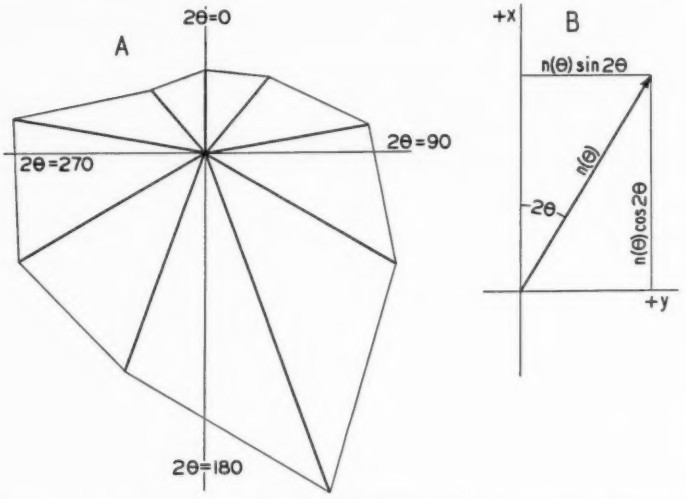

FIG. 7.—*A*, radius vector diagram of till pebble *a*-axes. *B*, method of resolving vector into components.

and these components are summed. The ratio of the sums of the horizontal to the vertical components yields the tangent of twice the mean azimuth of preferred orientation.[15]

The method of computing this mean for the *a*-axes of the till pebbles is shown in Table 3. The first column has the mid-point of the combined $\theta + (\theta + 180°)$ classes, and the third column shows the frequency $n(\theta)$ in each class. The second column lists 2θ, obtained by doubling the values in the first column. The fourth column lists sin 2θ, and the fifth column carries the product $n(\theta)$ sin 2θ, with the

[15] This method is really that used in finding the phase of the second harmonic by Fourier analysis.

sum below, having due regard for positive and negative values. The sixth and seventh show similar data for cos 2θ and $n(\theta)$ cos 2θ, also summed below.

Figure 7, *B*, shows the process graphically. Each radius vector is resolved into the components shown; the sum of the horizontal

TABLE 3

COMPUTATION OF MEAN PREFERRED ORIENTATION OF TILL PEBBLES BY RADIUS-VECTOR SUMMATION METHOD

Azimuth Classes	Mid-point θ	2θ	$n(\theta)$	sin 2θ	$n(\theta)$ sin 2θ	cos 2θ	$n(\theta)$ cos 2θ
350– 10 170–190	0°	0°	5	0.000	0.000	+1.000	+ 5.00
10– 30 190–210	20	40	6	+ .643	+ 3.86	+0.766	+ 4.60
30– 50 210–230	40	80	10	+ .985	+ 9.85	+0.174	+ 1.74
50– 70 230–250	60	120	13	+ .866	+11.30	−0.500	− 6.50
70– 90 250–270	80	160	22	+ .342	+ 7.52	−0.940	−20.70
90–110 270–290	100	200	14	− .342	− 4.80	−0.940	−13.20
110–130 290–310	120	240	13	− .866	−11.30	−0.500	− 6.50
130–150 310–330	140	280	12	− .985	−11.80	+0.174	+ 2.08
150–170 330–350	160	320	5	−0.643	− 3.22	+0.766	+ 3.83
Total					+ 1.41		−29.65

components is $\Sigma\,[n(\theta) \sin 2\theta]$, and the sum of the vertical components is $\Sigma\,[n(\theta) \cos 2\theta]$. The ratio of the horizontal to the vertical components yields the tangent of the angle $2\theta_m$; in terms of the summed values this tangent will be associated with the mean angle of azimuth, θ_m:

$$\tan 2\theta_m = \frac{\sum[n(\theta) \sin 2\theta]}{\sum[n(\theta) \cos 2\theta]}, \qquad (1)$$

from which $2\theta_m$ can be found from trigonometric tables. It should be mentioned that the choice of axes in Figure 7, *B*, was governed by the convention of measuring the azimuthal angle clockwise from 0°. This choice of axes permits the direct use of standard trigonometric tables for both positive and negative values of the functions.

Substitution of the data from Table 3 into equation (1) yields

$$\tan 2\theta_m = \frac{+1.41}{-29.65} = -0.047 .$$

The value −0.047 is found to be the tangent of −3° and −183°. (In periodic functions $\tan \theta = \tan (180° + \theta)$; the negative sign in the present case indicates that the angle must be measured counterclockwise around the azimuthal circle.) Inspection of Figure 7, *A*, indicates that the longer radius vectors are associated with the region $2\theta = 180°$, so that the proper choice is

$$2\theta_m = -183° ,$$

$$\theta_m = -91°.5 .$$

By measuring an angle of 91°.5 counterclockwise from 0° in Figure 3 (left), azimuth 268°.5 is indicated as the mean. This line is, of course, the same as 268°.5 − 180° = 88°.5. Thus the mean preferred orientation lies approximately east-west.

The radius-vector summation method is somewhat tedious, although the organization of Table 3 applies to all cases, with changes only for the frequencies in each class. The radius-vector method becomes more complicated if one wishes to find the average degree of spread of the data about the mean orientation. For this reason the writer prefers an empirical but sound method of organizing the data which permits the direct application of conventional moment analysis. By this method not only the mean orientation but the spread, the degree of asymmetry, and other characteristics of the azimuthal distribution are readily obtained.

II. MOMENT ANALYSIS OF AZIMUTHAL DATA

The data in columns one and three, Table 3, assembled as a histogram, yield Figure 8. This frequency distribution may be treated by

conventional methods to find the arithmetic mean azimuth, the standard deviation, and higher moments. Table 4 shows the "short method" of computing the first two moments. This method is described in every statistics text and consists simply in setting the origin of an arbitrary d-scale opposite the largest central class. The fifth column shows the product of the frequency times d, and the seventh column shows the product of the frequency times the square

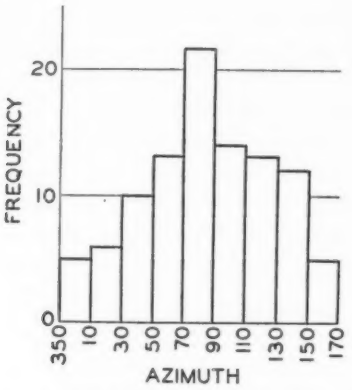

FIG. 8.—Azimuthal distribution of till pebble a-axes

TABLE 4

COMPUTATION OF MEAN ORIENTATION AND AVERAGE DEGREE OF SPREAD OF TILL PEBBLES BY MOMENT METHOD

Arbitrary Class	Mid-point	Frequency	d	fd	d^2	fd^2
350– 10	0°	5	−4	−20	+16	+ 80
10– 30	20	6	−3	−18	9	54
30– 50	40	10	−2	−20	4	40
50– 70	60	13	−1	−13	1	13
70– 90	80	22	0	0	0	0
90–110	100	14	+1	+14	1	14
110–130	120	13	+2	+26	4	52
130–150	140	12	+3	+36	9	108
150–170	160	5	+4	+20	+16	+ 80
Total	.....	100	.....	+25	.....	+441

of d. These columns are summed at the base of the table; the values divided by total frequency (100) yield $n_1 = 25/100 = 0.25$ and $n_2 = 441/100 = 4.41$. To find the arithmetic mean azimuth in degrees, n_1 is multiplied by the class interval (20°) and added to the mid-point of the $d = 0$ class (80°):

$$\text{Mean azimuth} = 80 + 20(0.25) = 85°0 .$$

This value represents a line drawn from 85° to 275° on Figure 3 (left). The radius-vector method yielded the value 88°.5, which is within 4 per cent of the present value.[16]

The average spread of the data in degrees about the mean azimuth (the standard deviation, σ) is found from the relation $\sigma = k\sqrt{n_2 - (n_1)^2}$, where k is the class interval, 20°:

$$\sigma = 20\sqrt{4.41 - (0.25)^2} = 41°6 .$$

The arithmetic mean azimuth found by this method is the first moment of the azimuthal distribution. It is located at the center of gravity of the distribution. The standard deviation is based on the second moment, and it measures the average spread of the data about the arithmetic mean. The value 41°.6 states that, on the average, the points lie within the range 85°.0 ± 41°.6 on the east, and 275° ± 41°.6 on the west. By sketching these lines on the left-hand diagram of Figure 3, it may be seen that about 70 per cent of the points fall within this range.

The preparation of the histogram of the a-axes was simplified by the coincidence of the modal class with the center of the histogram. In the general case this will not happen, and it is necessary to choose the origin of the histogram. The frequencies of the till-pebble c-axes, expressed in the same classes as the a-axes, yield the data in Table 5. Inspection of this table and of Figure 3 (right) shows that the preferred orientation lies approximately north-south, so that the large frequency associated with mid-point 0° should be at the center of the histogram. Inasmuch as there always are nine classes in this procedure, the modal class will have four classes on either side.

[16] In theory the two means should coincide, but the exact value is in part a function of the choice of class intervals and the sensitivity of the method.

Hence, in arranging these classes, the tabulation is divided into two parts at azimuth 90°, and the last four classes are transferred without altering their order to the head of the table, as shown in Table 6. These latter data are shown in Figure 9. The inset shows the figures in Table 5; note that the last four classes were simply translated to the left of the modal class. This method of arranging the histogram is based on the principle that the significant item is the distribution itself and not the arbitrary point of breaking the circle into classes. In essence it involves examining the distribution of points about the circle and building a histogram about the central modal class. In some instances there is so little preferred orientation that the method does not yield a significant figure. One such instance will be given later.

TABLE 5

Arbitrary Class	Frequency
350–10	30
10–30	16
30–50	11
50–70	5
70–90	3
90–110	4
110–130	7
130–150	8
150–170	16

TABLE 6

Arbitrary Class	Frequency
90–110	4
110–130	7
130–150	8
150–170	16
350–10	30
10–30	16
30–50	11
50–70	5
70–90	3

For the distribution of *c*-axes shown in Figure 9 the radius-vector method yields a mean azimuth of 0°.5, and the moment method yields a mean azimuth of 359°.5, which agrees within 1 per cent. The standard deviation of the distribution is 36°.2, indicating less spread about the mean than the *a*-axes show.

Once the principle of preparing these histograms is granted,[17] other statistical tests may be used. For example, Figures 8 and 9 are fairly symmetrical, suggesting distributions of the Gaussian type, although Figure 9 is more "peaked" than Figure 8. Tests for symmetry may be made with the third and fourth moments,[18] but a

[17] Azimuthal distributions represent no new departure, inasmuch as they have been used in physics to represent the scattering of protons, etc.

[18] Krumbein, "Size Frequency Distributions of Sediments and the Normal Phi Curve," *Jour. Sed. Pet.*, Vol. VIII (1938), pp. 84–90.

simple graphic test may be made with probability paper (Codex No. 3127). Probability paper is so designed that the cumulative curves of symmetrical distributions plot as straight lines. Figure 10 shows the cumulated data of Figures 8 and 9 plotted on probability paper. The agreement with straight lines is satisfactory and supports the conclusion that the distributions are essentially symmetrical. The significance of this test is its suggestion that, on the average, the probabilities are equal that a given point will fall on either side of the mean, essentially in accordance with the laws of chance.

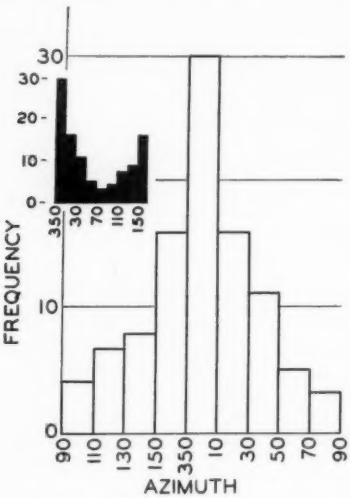

FIG. 9.—Azimuthal distribution of till pebble *c*-axes. Inset shows order in which data are drawn from original distribution (Table 5).

The cumulative curves of Figure 10 may be used to determine the arithmetic mean directly. This is done by reading the azimuth value corresponding to the 50 per cent line. For the *a*-axes the value is 86°; for the *c*-axes it is 359°, which agree well with the calculated values. The use of cumulative curves made in the conventional manner permits the development of a set of quartile measures to describe the statistical properties of the distribution. The median azimuth is read from the 50 per cent line, and the 25 and 75 per cent quartiles afford a basis for measures of spread and asymmetry.[19]

III. MOMENT ANALYSIS OF DIP ANGLES

To supplement the analysis of mean azimuth, a similar study of mean angle of dip may be made, independent of azimuth. To draw the dip distribution from the original data, the range 0° – 90° may

[19] Details of quartile methods may be found in Krumbein and Pettijohn, *op. cit.*, chaps. viii and ix.

be divided into 10° classes, each of which is bounded by concentric circles (Fig. 3). The number of points in each class is counted and assembled into a frequency table. Border-line cases may be divided between adjacent classes. Figure 11 shows the *a*-axes of the till pebbles arranged according to their angles of dip, using classes 0°–9°;

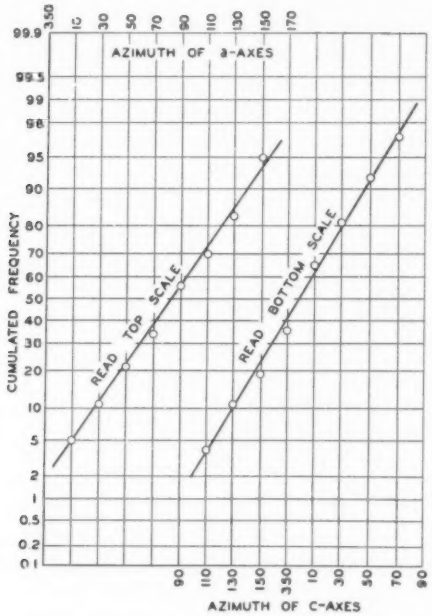

FIG. 10.—Probability graphs of cumulated azimuthal distributions of till pebble *a*- and *c*-axes.

10°–19°, etc. Most of the axes have low dips, which renders the figure asymmetrical. The arithmetic mean angle of dip is computed by the method of moments described previously; as before, higher moments may also be found if desired. The mean angle of dip of the data in Figure 11 is 24°.

The *c*-axes of the till pebbles show an entirely different picture. Figure 12 shows the distribution of dip angles. The arithmetic mean of this distribution is 48°. Inasmuch as the *c*-axes are the poles of

the maximum projection planes, the inclination of the planes is complementary to that of the axes. Hence, the mean inclination of the pebble planes is $90° - 47° = 43°$. This indicates that the pebbles, on the average, lie with their maximum planes tilted at nearly 45°. Thus, the pebbles do not lie "flat" in the till, nor do they stand "on edge"; the gentle inclination of the *a*-axes indicates that they ride an essentially horizontal "keel."

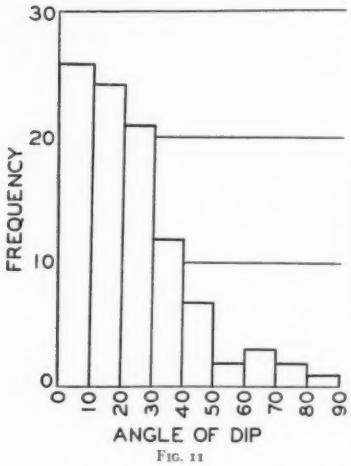

FIG. 11

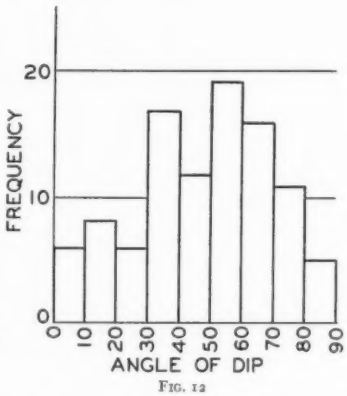

FIG. 12

FIG. 11.—Distribution of till pebble *a*-axes according to angle of dip

FIG. 12.—Distribution of till pebble *c*-axes according to angle of dip

APPLICATIONS OF STATISTICAL DATA

The statistical analysis of azimuth and dip independently of each other supplies data which not only supplement the petrofabric diagrams and the polar co-ordinate chart but, in addition, pave the way for further detailed studies. For example, the petrofabric diagram does not indicate whether the axes are distributed normally (in the statistical sense) about the mean azimuth, but the numerical data afford a simple test for this property, as was illustrated in Figure 10.

Numerical data also pave the way for comparisons of preferred orientation with other characteristics of sediments. For example, the question may be raised whether pebbles of particular shapes are

better oriented with respect to the mean than others. Inasmuch as not only the mean azimuth is known but also the average degree of spread (σ), the orientation of any pebble may be expressed in terms of its departure from the mean. This departure may be defined as d/σ, where d is the difference between the azimuth of the pebble axis and the mean azimuth, and σ is the standard deviation of the sample. The shape may be expressed in several manners. Elsewhere[20] the writer defined the "elongation" of a pebble in terms of the operational axes as $E = \sqrt{bc}/a$. This index measures approximately the extent to which pebbles are non-equidimensional; spheres have unit value, and the index decreases toward zero as the pebbles become noticeably cylindrical or disk shaped. A comparison of the index of till pebbles with d/σ showed that, in general, the more "elongated" pebbles were better oriented than less elongated pebbles. A similar study relating size of pebbles to degree of preferred orientation showed no significant differences over the range of sizes represented in the samples, which included pebbles having long axes from about 20 to 80 mm.

Another study which may be made with numerical orientation data concerns the effect of continued transportation on the degree to which the pebbles are oriented. This can readily be measured in terms of σ. A typical study may involve a deposit of glacial till with pebbles derived from a given ledge of limestone. Samples of pebbles may be collected at intervals away from the source ledge, and the degree of orientation measured. If there is a progressive decrease of σ as a function of distance, the inference is that the degree of orientation is improved as transportation continues. The writer has some preliminary data on this subject but not enough to establish the argument. Pebbles were collected within 50 feet of the source outcrop, and it was found that a substantial number were oriented transverse to the known direction of ice movement. The result was a large σ for the sample. A tentative explanation is that, in plucking fragments from bedrock, the ice acquires the fragments oriented with their maximum cross-sectional areas normal to the direction of ice movement. This orientation affords the best leverage, owing to the

[20] Krumbein, "Application of Photoelectric Cell . . . . ," *op. cit.*

large surface-weight ratio exposed. As these particles are transported, the ice tends to rotate them into coincidence with the direction of ice movement; and consequently it may be expected that σ will decrease with distance to some limiting value characteristic of ice transportation.

Another advantage of numerical data arises from their usefulness in studying the areal variation of preferred orientation. Richter,[21] for example, illustrated the direction of glacial movement by arrows pointing in the direction of preferred orientation. Reiche used a similar device in his study of cross-lamination.[22] An extension of such maps involves plotting the numerical value of σ and drawing contours to represent the variation in degree of orientation from place to place. The writer and Miss Aberdeen[23] referred to such maps as "environmental patterns" of sedimentary attributes; the orientation maps may be compared with similar maps of size, degree of sorting, shape, and the like, to learn how sedimentary characteristics vary from place to place within an environment of deposition.

The possible benefits which may accrue from the study of numerical orientation data do not eliminate the advantages of studying the petrofabric diagrams themselves. The two approaches properly supplement each other. Petrofabric diagrams present the picture as a whole and are useful in establishing the play of forces which control the orientation. Moreover, there are numerous instances in which the particles have neither a marked azimuthal concentration nor a marked dip distribution, and yet the two in combination yield a significant petrofabric pattern. To round out the discussion and to illustrate an instance of the type mentioned, the preferred orientation of two other deposits is included here.

ORIENTATION OF OUTWASH PEBBLES AND BEACH PEBBLES

I. OUTWASH PEBBLES

A sample of 100 pebbles was collected from a late Wisconsin outwash terrace along Fox River, near Cary, Illinois. The sample was

[21] K. Richter, "Gefüge und Zusammensetzung des norddeutschen Jungmoränengebietes," *Zeits. für Geschiebeforschung, Beihefte,* Vol. IX (1933), pp. 1–63.

[22] *Op. cit.*

[23] W. C. Krumbein and Esther Aberdeen, "The Sediments of Barataria Bay," *Jour. Sed. Pet.*, Vol. VII (1937), pp. 3–17.

taken from a single gravel bed which dipped 12° due south. The field relations suggest a deltaic deposit, although the dip of the beds seemed, on the whole, to be more gentle than may be expected. The orientation of the *a*- and *c*-axes was measured as before and yielded the data shown in Figure 13. The diagrams indicate that the long *a*-axes dip at comparatively low angles and are spread more widely along the periphery of the circle than are the axes of the till pebbles.

Figure 14 shows the petrofabric diagrams of the outwash pebbles. The diagrams are oriented with the north point at the top. The di-

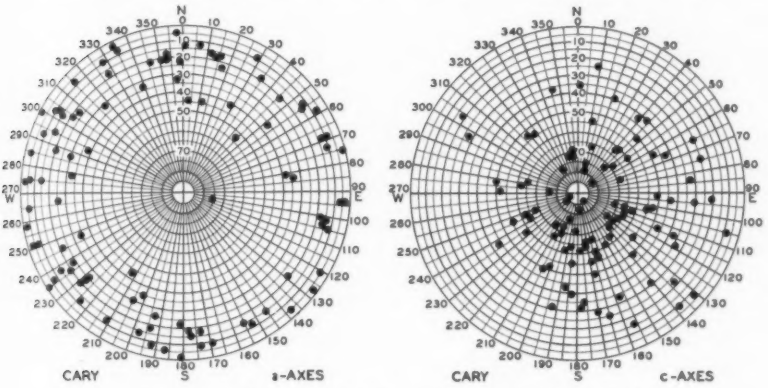

FIG. 13.—*Left*, distribution of *a*-axes of Wisconsin outwash pebbles. *Right*, distribution of *c*-axes of same sample.

rection of stream flow was from north to south, or from a point slightly west of north to slightly east of south. The monoclinic symmetry of the diagrams agrees substantially with this observation, inasmuch as the *a*-fabric axis (the axis of movement) lies approximately in the direction of flow.

Histograms of the *a*- and *c*-axes, prepared as previously described, are shown in Figure 15, *A* and *B*. In neither case is the mode conspicuous, and secondary modes are present. Interestingly enough, both the *a*- and *c*-axes show their modes at approximately the same azimuth. For the *a*-axes the moment method yields a mean of 359°.6; for the *c*-axes, 352°. This is an instance in which the numerical data are weak because of the absence of a pronounced modal class. Even the radius-vector method is of little aid, because a dis-

tribution with no preferred orientation whatever always yields the value 0°. The essential coincidence of a poor mode with 0° in these cases throws some doubt on the values. The very weakness of the orientation is reflected by the standard deviations of the two distributions, which are 52°.4 for the *a*-axes and 48°.6 for the *c*-axes. These values are noticeably larger than those of the till pebbles, which were 41°.6 and 36°.2, respectively.

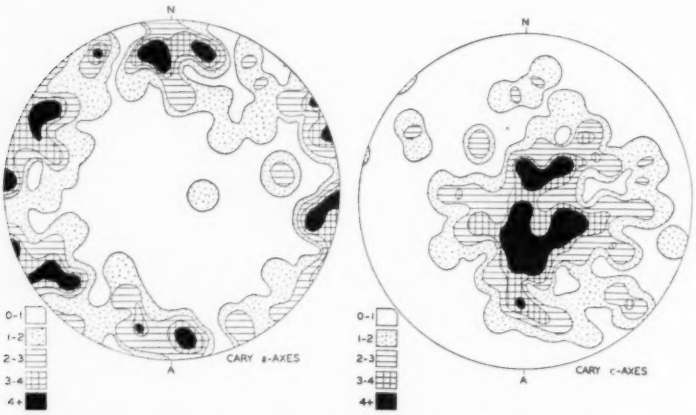

FIG. 14.—Petrofabric diagrams of outwash pebbles. Data from Fig. 13

Wadell[24] used histograms of dip direction which extended over the entire 360°, with the data assembled in 40° classes. Such a diagram of the outwash *c*-axes would yield a fairly pronounced mode at 170°; but the center of gravity occurs at 152°, because of a high degree of asymmetry in the distribution. Despite the weakness of the several statistical values the data do converge toward a direction of flow slightly east of north-south. The lack of critical patterns in such data was noted by Richter,[25] who observed a wide spread of the data. Wadell, however, found convincing evidence of preferred orientation in delta pebbles and noted that the long axes were imbricated upstream. The Cary outwash pebbles agree with this aspect of Wadell's work.

[24] *Op. cit.* An example is given in Krumbein and Pettijohn, *op. cit.*, p. 269.

[25] "Ergebnisse und Aussichten . . . . ," *op. cit.*

Insets *C* and *D* of Figure 15 show the dip-angle distribution of the *a*- and *c*-axes. The *c*-axes are noticeably steeper than the *a*-axes; the mean angle for the former is 54°, and for the latter is 18°. The 54° dip of the *c*-axes indicates that the average inclination of the maximum projection planes is about 36°, so that, on the whole, the pebbles are more "flat-lying" than in the till. A tentative interpre-

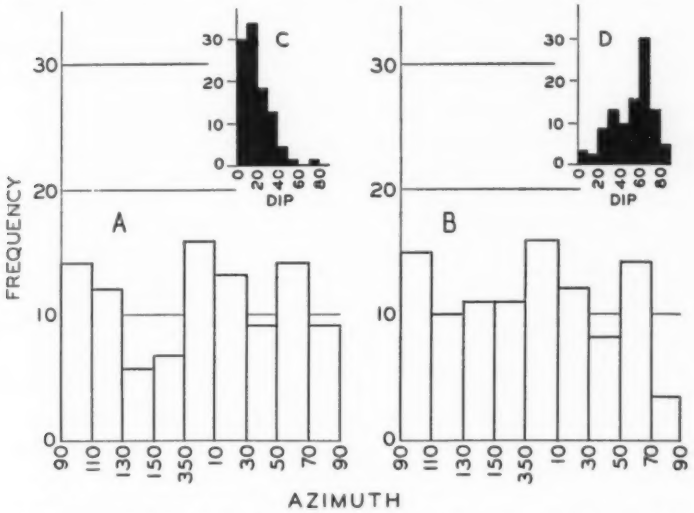

FIG. 15.—*A*, azimuthal distribution of outwash pebble *a*-axes. *B*, azimuthal distribution of *c*-axes. Insets show *C*, distribution of dip angles of *a*-axes; *D*, distribution of dip angles of *c*-axes.

tation of these data seems to be that the rapid deposition of the pebbles from the outwash waters resulted in a wide scatter of the pebble axes about the underlying imbrication pattern, perhaps owing to shape variations in the pebbles and velocity variations in the transporting agent. The latter factor may mix pebbles from traction with those from saltation, resulting in the superposition of two patterns. Richter, in this connection, concluded that the smaller pebbles lined up with the direction of flow, and the larger ones transverse to it.[26]

[26] Knopf and Ingerson, *op. cit.*, p. 203.

This is an instance in which the petrofabric diagrams yield a more significant picture of the data than do the azimuth or dip distributions. In terms of an areal study, however, it is possible that the degree of imbrication may improve in the direction of transportation, and the numerical value of σ plotted against distance could be used to determine the extent of such improvement.

II. BEACH PEBBLES

One of the most striking sets of oriented pebbles studied by the writer was collected from the beach at Little Sister Bay, Wisconsin. The beach is composed almost wholly of disk-shaped dolomite pebbles.[27] Detailed study showed a progressive imbrication of the pebbles away from the water. Near the water's edge the pebbles lay nearly flat, and within a few feet of the water line the imbrication became very pronounced, with the dips of the pebbles increasing until they stood nearly on edge. Landward from this steepest portion the pebbles lay flatter again, with another increase in steepness as the line was followed farther inland. The sample was collected from one of the steepest portions of the series. The leveling frame was held horizontally above the pebbles in marking them, and in the laboratory the pebbles were mounted in the goniometer by viewing the reference lines from a point above the vertical circle.

The data obtained are shown in Figures 16 and 17. The long *a*-axes are distributed over roughly half a circle, and with few exceptions dip lakeward or along the shoreline, which strikes N. 80° E. The *c*-axes, on the other hand, are concentrated in the southern half of the diagram and dip fairly steeply. The petrofabric diagrams (Fig. 17, *A* and *B*) are arranged with the shoreline (*AA*) horizontal and the lakeward direction toward the north. The two diagrams have monoclinic symmetry with the *a*-fabric axis (axis of movement) normal to the line *AA*.

A study of the azimuthal distribution of these axes brings out an important point in the interpretation of axial trends. The petrofabric diagram of the *a*-axes suggests an approximately equal distribution of axes over the lakeward half of the diagram, but actually

[27] W. C. Krumbein and J. S. Griffith, "Beach Environment in Little Sister Bay, Wisconsin," *Bull. Geol. Soc. Amer.*, Vol. XLIX (1938), pp. 629–52.

there are more axes parallel to the shoreline than in any other particular direction. This is brought out by the histogram of these axes in Figure 18, *A*. The mean azimuth by the moment method is 76°, very nearly parallel to the strike of the shore. This concentration in the modal class arises from the fact that azimuthal classes with mid-points from 40° to 100° receive contributions from their counterparts on the opposite side of the circle, whereas mid-points 300°–340° and 0°–30° receive no contributions from the opposite classes.

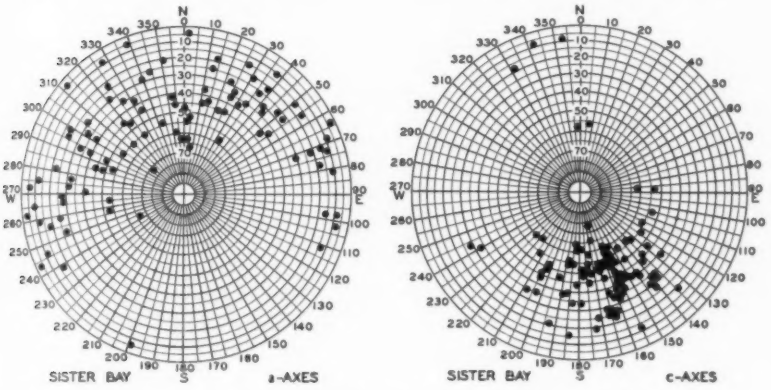

FIG. 16.—*Left*, distribution of *a*-axes of beach pebbles. *Right*, distribution of *c*-axes of same sample.

Thus the azimuth distribution shows a fairly pronounced mode at 80°, although the rest of the distribution is approximately equally distributed over the range. In this instance the azimuthal distribution emphasizes an important trend which may have been overlooked in the petrofabric diagram.

The azimuthal distribution of *c*-axes is shown in Figure 18, *B*. The mean azimuth is 163°.2, which is very nearly normal to the shoreline. The standard deviation of the *a*-axes is 48°.4, a fairly large value consistent with the nature of the distribution; for the *c*-axes, $\sigma = 27°.8$, indicating a markedly compact distribution.

The petrofabric diagram of the beach pebble *c*-axes (Fig. 17, right), suggests a method of reducing such diagrams to a single point of mean concentration. The mean azimuth is 163°.2, and the

mean angle of dip (Fig. 18, *D*) is 44°. A single point drawn on a circle with these co-ordinates indicates the mean center of polar orientation; a plane drawn at 90° to this pole would indicate the mean orientation of the maximum projection planes. It would be advantageous if all types of diagrams could be summarized in such a manner, but frequently the symmetry is such that it can only be done in terms of half-circles. It may be desirable to consider each half-circle independently, with reference to some principal axis, as the *a*-fabric axis or its normal in the horizontal plane.

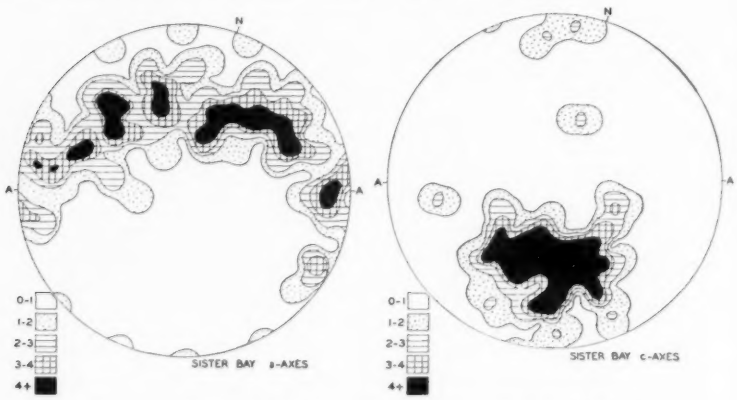

FIG. 17.—Petrofabric diagrams of beach pebbles. Data from Fig. 16

The arrangement of the beach pebble *a*-axes parallel to the shoreline agrees with the discussion of beach imbrication by Fraser,[28] who pointed out that disk-shaped pebbles are transported with their longest axes normal to the direction of wave motion. Upon deposition the axes tend to remain in this orientation. Fraser also remarked that the maximum planes of the pebbles usually lie nearly flat or are inclined toward the water. The *c*-axes in this sample show an average dip of 44° away from the water (Fig. 18, *D*), which means that the maximum planes incline toward the water at about 46°. Elsewhere on the beach, however, the pebbles are less steeply inclined. The mean dip of the *a*-axes (Fig. 18, *C*) is 31°. This dip is

[28] H. J. Fraser, "Experimental Study of the Porosity and Permeability of Clastic Sediments," *Jour. Geol.*, Vol. XLIII (1935), pp. 910–1010.

larger than that implied by Fraser for the long axes; but it is consistent with the steep inclination of the maximum planes, which permits a wider fluctuation in the dip angle of the *a*-axes than more flat-lying pebbles would permit. The generally striking agreement

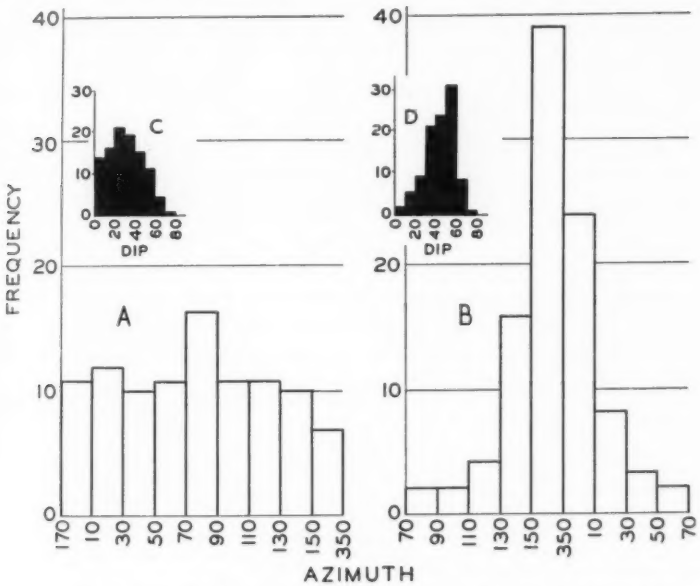

FIG. 18.—*A*, azimuthal distribution of *a*-axes of beach pebbles. *B*, azimuthal distribution of *c*-axes. Insets show *C*, distribution of dip angles of *a*-axes; *D*, distribution of dip angles of *c*-axes.

of theory and observation here is probably due to the unusual uniformity of pebble shapes on the beach. The absence of many roller pebbles permits a clean-cut imbrication of the disks.

GENERAL SUMMARY

Detailed discussion of single samples of oriented pebbles is justified mainly to the extent that they serve as examples of lines of attack which are available for such data. The writer is of the opinion that single samples are of limited value for genetic studies because of variations which may accompany particular environmental con-

ditions. The main purpose of the paper is to describe methods and to show that conventional statistical methods may readily be used with orientation data. The radius-vector method was taken directly from harmonic analysis, and the moment method is described in every statistics textbook. Regardless of the pattern of the orientation diagram, the underlying statistical principles are the same, but varying degrees of preferred orientation endow the numerical data with greater or lesser significance. As in all statistical work, the judgment of the worker plays a role.

If there is an implied criticism of petrofabric diagrams in this paper, it is that of sole reliance on those diagrams, without recourse to other methods. In the writer's opinion, the two methods supplement each other. The petrofabric diagram affords a single composite picture of the fabric, whereas the statistical data provide numerical measures for testing similarities and differences among samples from the same or different deposits. As orientation analysis of sediments proceeds, the relative roles to be played by various devices will obviously depend upon their value in interpreting the data analytically.

ACKNOWLEDGMENTS.—The writer is indebted to various workers for helpful criticisms and suggestions, especially for the role played by the petrofabric diagrams themselves. Drs. Eleanora B. Knopf, James Gilluly, and Earl Ingerson were particularly helpful in this connection. Dr. Carl Eckart, of the department of physics of the University of Chicago, gave helpful advice on the treatment of the azimuthal distributions. Mr. J. Scott Griffith collaborated in the collection of the oriented samples.

LATE QUATERNARY UPWARPINGS OF NORTHEASTERN NORTH AMERICA

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ABSTRACT

The crustal upwarpings upon decrease and removal of the load of the last Labradoran ice sheet appear to have occurred spasmodically and intermittently during ages of several hundred to a few thousand years' duration, separated by somewhat longer ages of quiescence. The most important uplifts took place prior to, and during, the first stages of the glacial lakes Algonquin and Iroquois, during the Champlain Sea age, and during the early and the late Postglacial. With the exception of that during the Nipissing age, the long quiescences were caused by increase of the ice or by a balance between supply and wastage; and, when established, they prevailed until the accumulating stresses had increased to the yielding point. The most important quiescences were those during the deglaciation of New England (perhaps regionally *equal* uplift); during the three-outlet stage of Lake Algonquin, Lake Iroquois proper, and Lakes Vermont and Frontenac; during the Ottawa Sea (possibly some land sinking); and during the Nipissing Great Lakes.

Even with quiescence during the uncovering of New England, the contemporaneous retreat of the shoreline, indicated by the weak development of all elevated marine shorelines, can have been produced by a lowering of the sea-level, for the Patrician, Keewatin, and Cordilleran ice sheets were still growing. The Algonquin and the Nipissing quiescences are well proved by exceptionally strong beaches connecting respectively three and two outlets. The Ottawa Sea, whose shore stands at 240 feet altitude at Ottawa, is now believed to be the water body in which the upper marine clay of that region was deposited. The stage was established by a transgression of about 50 feet, produced by rise of the sea-level during quiescence and possibly some sinking of the land. The Ottawa Sea climaxed during the transition from the late Glacial to the Postglacial. The upwarpings appear to have proceeded in accord with R. A. Daly's hypothesis of flow at *great* depth.

INTRODUCTION

The study of the late Quaternary upwarpings of northeastern North America has made notable progress during the last fifteen years. Daly's[1] punching hypothesis of the vertical movements of the earth's crust upon accumulation and unloading of the ice sheets, implying flow at *great* depth, has appeared and may replace the wave hypothesis, which assumed flow at *small* depth. Levels of successive glacial lakes in the Connecticut Valley have proved to be parallel,[2] and so have the water planes of glacial Lake Vermont and the subse-

[1] R. A. Daly, *The Changing World of the Ice Age* (New Haven: Yale University Press, 1934), p. 122.

[2] R. J. Lougee, "Hanover Submerged," *Dartmouth Alumni Mag.*, May, 1935, p. 6.

quent Champlain Sea.[3] Glacial Lake Algonquin has been found to have terminated not by pouring as a result of tilting but by lowerings as successively lower outlets were uncovered by the retreating ice.[4] By utilizing these and other new ideas, observations, and correlations, an attempt will be made to present a new interpretation of the movements of the shoreline at Ottawa, Canada, to review the main events of the upwarpings in a larger area, and to correlate them with stages of the Great Lakes, sections of the Niagara Gorge, and a tentative chronology of the late Glacial and the Postglacial.

The writer is indebted to Professor R. A. Daly and to Mr. W. A. Johnston for good advice.

LATE GLACIAL UPWARPINGS OF THE NORTHEASTERN STATES AND THE GREAT LAKES REGION

The thickness of the water layer removed from the oceans during the last Wisconsin-Würm glacial climax was probably about 300 feet.[5] To be sure, the extent of the ice in Asia is now known to have been larger than was held when the estimate was made;[6] but, on the other hand, the glacial culminations are believed to have occurred at somewhat different times in western and eastern North America,[7] and, by inference, also in Europe and Asia. Since the ocean floor may have risen in response to the decreased load about one-fifth of the thickness of the removed water,[8] the actual lowering of the ocean surface was perhaps roughly 240 feet.

Conversely, the added weight of the thick ice sheets pressed down the earth's crust in the areas of continental glaciation to amounts

[3] D. H. Chapman; "Lateglacial and Postglacial History of the Champlain Valley," *Amer. Jour. Sci.*, Vol. XXXIV (1937), pp. 89–124; see esp. pp. 118 and 122.

[4] G. M. Stanley, "Lower Algonquin Beaches of Penetanguishene Peninsula," *Bull. Geol. Soc. Amer.*, Vol. XLVII (1936), pp. 1933–60; see esp. pp. 1951 and 1958; "Lower Algonquin Beaches of Cape Rich, Georgian Bay," *ibid.*, Vol. XLVIII (1937), pp. 1665–86; see esp. pp. 1680 and 1685.

[5] Ernst Antevs, "The Last Glaciation," *Amer. Geog. Soc. Res. Ser. 17* (1928), p. 81.

[6] V. A. Obruchev, "Die Verbreitung der Eiszeitspuren in Nord- und Zentralasien," *Geol. Rundschau*, Vol. XXI (1930), pp. 243–83; P. D. Krynine, "Pleistocene Glaciation of Siberia," *Amer. Jour. Sci.*, Vol. XXXIV (1937), pp. 389–400.

[7] Antevs, "Climatic Variations during the Last Glaciation in North America," *Bull. Amer. Met. Soc.*, Vol. XIX (1938), pp. 172–76.

[8] Daly, *op. cit.*, pp. 115, 138, and 147.

which increased toward the centers. The crustal uplift, upon decrease and removal of the ice load, has consisted of a centripetally increasing amount prior to the recording of the marine limit, the amount indicated by the present altitude of the highest late Glacial marine shoreline plus an amount equal to the rise of the sea-level since the marine limit was registered. The marine limit indicates the excess of the rise of the land above that of the sea since the uncovering. The zero isobase marks points with equal rise of the land and of the sea since their release from the ice. And, if the actual maximal lowering of the sea-level was 240 feet, the isobase of −240 feet, which runs some 140 miles south of Boston and 75 miles south of New York,[9] indicates the approximate periphery of the raised land. Farther west the limit of the depressed and later elevated land is possibly marked by the Whittlesey hinge line, which passes from the northeast corner of Ohio, through Lake St. Clair to Milwaukee.[10] The marine limit rises from zero at a point 20 miles south of Boston to 200 feet on the coast of Maine, 300 feet on a line 15–30 miles inside the coast of Maine, and 717 feet in Montreal.

As to events, some differential uplift took place during the release from the last ice sheet of the Plainville-Tariffville belt in northern Connecticut.[11] Because the water planes of glacial Lake Hitchcock and the subsequent Lake Upham in the Connecticut Valley are parallel,[12] Lougee believes that during the ice retreat from Tariffville to St. Johnsbury and Littleton in northern New England, or for more than 4,000 years, the land remained stable. At first thought, this conclusion may seem to be refuted by the weak development of all elevated marine shores in New England,[13] which show that re-

[9] Antevs, "The Last Glaciation," *op. cit.*, p. 85.

[10] F. B. Taylor in Frank Leverett and F. B. Taylor, "The Pleistocene of Indiana and Michigan and the History of the Great Lakes," *U.S. Geol. Surv. Mono. 53* (1915), pp. 503 and 506.

[11] Lougee, "Physiography of the Quinnipiac-Farmington Lowland in Connecticut," *Colby Mono. 7* (1938), pp. 55 and 61.

[12] Lougee, "Hanover Submerged," *op. cit.*, p. 6.

[13] Antevs, "Late Quaternary Changes of Level in Maine," *Amer. Jour. Sci.*, Vol. XV (1928), pp. 319–36; see pp. 329 and 334; N. G. Hörner, "Late Glacial Marine Limit in Massachusetts," *Amer. Jour. Sci.*, Vol. XVII (1929), pp. 123–45; see esp. p. 139; I. B. Crosby and R. J. Lougee, "Glacial Marginal Shores and the Marine Limit in Massachusetts," *Bull. Geol. Soc. Amer.*, Vol. XLV (1934), pp. 441–62; see esp. p. 447.

gression, broken at most by brief halts, took place from the time of deglaciation until the coast line had moved outside the modern shore, which is one of transgression. Yet, even with quiescence of the land during the uncovering of New England, regression of the shoreline could perhaps have taken place by lowering of the sea-level, for, while the Labrador ice was shrinking, the Patrician, Keewatin, and Cordilleran ice sheets were still growing. However, the parallelism of the lake planes could also have been produced by regionally *equal* uplift during the stated time, just as equal uplift occurred during the uncovering of northwestern Finland, according to the detailed studies by Sauramo.[14] What actually took place remains to be determined. Quiescence of the land or equal uplift, the latter implying retardation in the north, could possibly have been caused by the contemporary growth of the ice west of a line approximately from Chicago to the south end of James Bay.

At any event, differential uplift set in when the ice margin was at Littleton and St. Johnsbury,[15] and probably in the region of Schuylerville, 30 miles north of Albany. Undoubtedly the same upwarping began in the Great Lakes region as the ice withdrew from the Port Huron moraines,[16] standing just south of Lake Huron. The uplift coincided with retreat along the ice front from the Atlantic to Minnesota. The tilting continued until the outlet of Lake Algonquin at Fenelon Falls, 70 miles north-northeast of Toronto, had been raised to the level of the Port Huron and Chicago outlets,[17]

[14] Matti Sauramo, "The Mode of the Land Upheaval in Fennoscandia during Late-Quaternary Time," *Compt. Rend. Soc. Géol. de Finlande*, No. 13 (1939); see pp. 10, 12, 13, and 16.

[15] Lougee, "Hanover Submerged," *op. cit.*, p. 6.

[16] Taylor, *op. cit.*, pp. 503, 504, and 506.

[17] The Fenelon Falls (Kirkfield) stage of Lake Algonquin may have been brief. The chief reasons why Taylor ("Niagara Falls and Gorge," *12th Internat. Geol. Cong. Guide Book No. 4* [1913], pp. 5–70; see esp. p. 29; "The Pleistocene of Indiana. . . . ," *op. cit.*, pp. 410 and 413) believed it to be of long duration were the erosional effects of the outlet river (the Algonquin River) and the long time it must have taken to erode the Old Narrow gorge of the Niagara Canyon, the supposed correlative. However, most of this erosion may rather have occurred during the subsequent three-outlet phase. The abundance of worn beach shingle which Stanley ("Impounded Early Algonquin Beaches at Sucker Creek, Grey County, Ontario," *Papers of Mich. Acad. Sci., Arts, and Letters*, Vol. XXIII [1937], pp. 477–95; see esp. p. 487) presents as evidence for a long duration of the stage can be partly or largely due to ready access to glacial outwash and to the

and until Lake Iroquois proper had been established by rise of the water level south of the pivot line running from the outlet at Rome, New York, to Cobourg, Ontario.

Then there followed a long age of quiescence, as shown by the formation of the strong upper Algonquin beach while all the three outlets were in use[18] and by the development of the contemporaneous and equally strong Iroquois shore.[19] Regionally equal uplift is not probable in this case, because the southern part of the Algonquin beach is horizontal, implying that it had about reached its modern vertical stand. The quiescence prevailed through the existence of Lake Iroquois and the succeeding Lake Frontenac into the Champlain Sea age. This is shown by the fact that the Algonquin River, which ceased to flow as soon as upwarping was resumed, cut a strongly marked channel to and below the level of Lake Ontario at Trenton,[20] which it could do first during the Champlain Sea stage. This, in turn, can be calculated from the height of the Iroquois shore and from the subsequent lake lowerings. Since the altitude of the Iroquois beach near Trenton is 632 feet,[21] and Lakes Iroquois and Frontenac were lowered about 280 and 226 feet, respectively, Lake Frontenac stood at Trenton about 352 feet and the Champlain Sea about 126 feet above the modern sea-level. Lake Ontario standing at 246 feet, the submerged part of the Algonquin channel is of Champlain Sea age. Some beaches on the north side of the basin, attributed by Coleman[22] to the Champlain Sea, may belong to Lake Frontenac.

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continual rise of the lake during the formation of the beaches. The beaches are only known in an area half a mile across. Equally long duration of the Fenelon Falls and the three-outlet stages of Lake Algonquin is out of accord with a brief early Iroquois stage and a long duration of Lake Iroquois proper, and the history of these stages in the Huron and Ontario basins was parallel.

[18] Taylor, "The Pleistocene of Indiana . . . . ," *op. cit.*, pp. 411 and 413.

[19] A. P. Coleman, "The Pleistocene of the Toronto Region," *Annual Rept. Ont. Dept. Mines*, Vol. XLI, Part VII (1932), p. 35.

[20] W. A. Johnston, "The Trent Valley Outlet of Lake Algonquin and the Deformation of the Algonquin Water Plane in Lake Simcoe District, Ontario," *Geol. Surv. Canada Mus. Bull. No. 23* (1916), p. 14; Coleman, "Glacial and Postglacial Lakes in Ontario," *Univ. Toronto Studies, Biol. Ser. No. 21* (1922), pp. 31 and 51.

[21] Johnston, *op. cit.*, map.

[22] "Glacial and Postglacial Lakes . . . . ," *op. cit.*, pp. 51 and 54.

In the Champlain basin the parallelism between the water planes suggests, in good agreement with the events and conditions in the Great Lakes basins, that quiescence prevailed from the uncovering of the Fort Ann region south of Lake Champlain, through the life of glacial Lake Vermont, to and including the inauguration of the Champlain Sea.[23]

During the life of Lake Iroquois the quiescence was perhaps largely caused by an approximate balance between the supply and wastage of the ice. Thus the ice border damming Lake Iroquois seems to have been almost stationary, for the Iroquois beach terminates at Havelock near the ice border through Stony Lake, which first permitted Lake Algonquin to drain past Fenelon Falls.[24] The contemporary recession of about 100 miles in the Champlain Valley was no doubt partly due to decreased supply of ice east of the Adirondacks and Montreal, as suggested by the subsequent breaking-up of the ice in the lower St. Lawrence Valley. However, quiescence prevailed through Lake Frontenac in spite of a moderate rate of ice retreat.

LATE GLACIAL MOVEMENTS OF THE SHORELINE AT OTTAWA

In 1924 the two late Quaternary marine clays at Ottawa, Canada, were held to be deposited in the Champlain Sea, and the beach at 240 feet altitude was believed to record a lake largely of Postglacial age.[25] Although observation by Professor J. W. Goldthwait and the writer in 1925 suggested that the 240-foot beach was formed by a marine estuary, the significance of this—that the upper marine clay was deposited in this estuary, the Ottawa Sea—was not then realized by the writer. Therefore, the writer's old normal section at Ottawa is incorrect, there being in reality two separate sets of top beds—namely, one above, and another below, the 240-foot level. All the

[23] Chapman, *op. cit.*, pp. 97, 101, 116, 118, and 122.

[24] Johnston, *op. cit.*, map; Coleman, "The Pleistocene of the Toronto Region," *op. cit.*, p. 45.

[25] Johnston, "Late Pleistocene Oscillations of Sea Level in the Ottawa Valley," *Geol. Surv. Canada Mus. Bull. 24* (1916), pp. 7 and 10; "Pleistocene and Recent Deposits in the Vicinity of Ottawa," *Geol. Surv. Canada Mem. 101* (1917), pp. 18 and 21; Antevs, "Retreat of the Last Ice Sheet in Eastern Canada," *Geol. Surv. Canada Mem. 146* (1925), pp. 67, 73, and 78.

observations in the Ottawa region south of the river[26] may be comprised in the following normal sections.

The section *above* the 240-foot beach at Ottawa—Profile I—has only one marine clay, the lower or Champlain clay. It is entirely of late Glacial age.

PROFILE I

g) At top, 28–45 feet cross-bedded, coarse sand with layers of fine gravel. The bed contains a great abundance of shells of *Macoma* sp., *Mytilus edulis*, *Saxicava rugosa*, and *Balanus crenatus*

f) 0–25 feet fine, loose, evenly stratified sand, poor in shells

Distinct limit

e) 10–55 feet clayey sand and sandy clay, of which particularly the upper part is rich in shells. Species found: *Portlandia arctica*, *S. rugosa*, *Macoma calcaria*, *M. baltica* var. *grönlandica*, *Astarte compressa*, *Nucula tenuis*, *Axinus flexuosus*, *Natica affinis*, *Neptunea despecta*, *Cylichna alba*, and *B. crenatus*.

Abrupt limit, but no evidence of erosion

d) 2–2½ feet stiff marine clay—the lower or Champlain clay—with *P. arctica*

c) 1–3 feet stiff, rather distinctly to indistinctly varved clay without shells, probably deposited in brackish water

b) 9–12 feet varved glacial fresh-water clay, silt, and sand

a) 0–18 feet till

Bedrock

The section *below* the 240-foot beach—Profile II—has two marine clays, viz., the lower or Champlain clay, and the upper or Ottawa clay. The top bed was about completed when the late Glacial came to a close.

PROFILE II

At top, 5–21 feet stiff marine clay—the upper or Ottawa clay—in the lower part containing shells, mostly of *P. arctica* and *S. rugosa*

Mostly abrupt limit

Beds *a* to *e* as in Profile I, preceding

Similar profiles were later observed west and north of Ottawa as far as Pembroke.[27] Of these sections, those named *A*, Blanche River, and *D*, Quyon, stand below the level of the Ottawa Sea and support the foregoing normal Profile II. Section *E*, Forester Falls, at about 430 feet, probably stands too high but is explainable by a drainage. Section *G*, Pembroke, with an intercalated erosion sur-

[26] Antevs, "Retreat . . . . in Eastern Canada," *op. cit.*, pp. 67–69 and 95–97.

[27] Antevs, "The Last Glaciation," *op. cit.*, pp. 97–101.

face at about 380 feet altitude, is subject to the same explanation as is *F*, Pembroke. The sections *B*, *C*, and *F* deserve special mention.

B) Kirks Ferry, 300 feet upstream on Blackburn Brook 10 miles north-northwest of Ottawa. Elevation of surface roughly 300 feet.

At top, 50 feet marine clay without shells
Marked limit
10 feet clay and sand. Shells of *Saxicava* abundant, of *Macoma* rare
10 feet covered
3 feet of stiff marine clay with *P. arctica*
Brook level

C) Brennan, roadcut 400 yards north of station, 28 miles north-northwest of Ottawa. Elevation of surface about 380 feet.

Several feet of clay, massive or with faint varves. Near base several shells of *P. arctica* and one shell of *Macoma*
Sharp limit
0–3 feet gravel with cobbles
Erosion surface
Several feet of sand and clay in alternating layers but not in varves. No shells observed. Depth to bottom unknown

F) Pembroke, brickyard at south edge of town. Elevation of surface, roughly 400 feet.

7 feet indistinctly varved clay, about 350 varves
1½ feet sand
Disconformity
⅓ foot leached clay
2 feet clay with fairly distinct varvity, 65 varves
1 foot gray-brown, rather stiff, nearly massive clay. Depth to bottom unknown

These sections show two marine clays, as do those below the 240-foot level near Ottawa. The clays in the two sets of profiles are probably correlatives. Thus the Ottawa Sea may have reached the altitude of nearly 400 feet at Brennan and at least 400 feet at Pembroke. The old shoreline may rise northward at a rate of about 5 feet per mile. The high rate of tilt may be explainable by the fact that the beach crosses the faults between the Paleozoic lowland and the pre-Cambrian upland.

The varved clay, bed *b* in the Ottawa profiles I and II, was deposited in glacial Lake Frontenac. Beds *c* and *d* record the introductory brackish-water stage and the developed Champlain Sea. This

marine stage was established by the lowering of the combined Lakes Frontenac and Vermont 226 feet to the sea across and through the thinned and decayed ice in the lower St. Lawrence Valley.[28] At the time, the ice border stood about 10 miles south of Montreal, on the north bank of the Ottawa River from Oka to Ottawa, and perhaps at Carleton Place and Whitney. The marine gulf covered the St. Lawrence–Champlain–Ottawa lowland and the Ontario basin. Marine mollusks spread as far as Brockville on the St. Lawrence; Fort Coulonge,[29] 60 miles west of Ottawa; and Venosta, 34 miles north of Ottawa. Near Venosta shells occur at 570 feet altitude. The highest limit of the Champlain Sea stands at 690 feet at Kingsmere, 8 miles northwest of Ottawa.[30] Glaciofluvial deltas attain elevations of about 600 feet at Kazubazua, 40 miles north of Ottawa, of 565 feet 17 miles southeast of Pembroke, and of 500 feet at Pembroke and 15 miles northwest of Pembroke.

As stated, quiescence prevailed at Fenelon Falls into the Champlain Sea age. In the Champlain basin differential, intermittent land uplift set in after the recording of the highest limit of the Champlain Sea.[31] In the region west-northwest of Ottawa, upwarping occurred just prior to the release from the ice, judging from the northward decrease in altitude of the glacial outwash plains. At Pembroke and Petawawa the pre-uncovering uplift may have amounted to more than 200 feet, for the glaciofluvial deltas stand at about 500 feet altitude, whereas the uplift of Pembroke may have been somewhat greater than that of Ottawa, the direction of the greatest tilting probably being about due north, as it was during the Ottawa Sea.

During the uplift the Champlain Sea deposits were eroded at Pembroke (*F* and *G*), Brennan (*C*), and Kirks Ferry (*B*). The sand and gravel beds at the localities *G*, *C*, and *B* were deposited either then or during the later transgression. When the waves broke

[28] Antevs, "Retreat . . . . in Eastern Canada," *op. cit.*, p. 65; Chapman, *op. cit.*, pp. 113 and 119.

[29] Winifred Goldring, "The Champlain Sea," *New York State Mus. Bulls. 239 and 240* (1920–21), pp. 153–94; see esp. p. 165.

[30] Johnston, "Pleistocene and Recent Deposits . . . . ," *op. cit.*, p. 17.

[31] Chapman, *op. cit.*, pp. 97, 101, 116, 118, and 122.

against shoals and islands of sand and gravel, now standing between the 300- and 400-foot contours on the Rideau River 6–7 miles south of Ottawa,[32] it washed out sand, bed *f* in the Ottawa Profile I. Since the base of this sand at the sand pits on the Rideau stands at roughly 270 feet and the ground surface at 340 feet, the sea-level at the time was perhaps in the vicinity of the 375-foot level, while the mouth of the Rideau was at Smith Falls. As the land continued rising and the water grew shallower, the cross-bedded gravelly top sand, bed *g*, was formed by drift from the near-by shore.

The regression of the Champlain Sea proceeded to below the level of the later Ottawa Sea, which was the result of a transgression. At locality 2, Billings Bridge, just south of Ottawa,[33] where clay was deposited to within some 15 feet of the surface of the Ottawa Sea, clayey sand at the 175-foot level suggests that the shoreline withdrew to about the 185-foot level, 50 feet below the Ottawa beach at about 235 feet. Section *B*, Kirks Ferry, described above, by its 50 feet of clay and sandy clay overlying 10 feet of clay and sand suggests that the subsequent transgression was more than 50 feet. Thus the Champlain Sea seems to have withdrawn to some 50 feet below the surface of the later Ottawa Sea, or about to the 190-foot level at Ottawa and the 350-foot level at Pembroke. At Ottawa the regression, which began at the altitude of 690 feet, amounted to about 500 feet (150 meters).

The uplift was started by the unloading of ice that took place during the life of Lake Frontenac and by the lowering of Lakes Iroquois and Frontenac-Vermont; and it was kept up by continued ice-melting and the drainage of Lake Algonquin. It may have been rapid. The regression of the shore took perhaps about as long as did that of the first 150 meters of the total 235 in northern Sweden, which consumed 3,200 years, averaging a foot in 6.4 years.[34] The upwarping probably occurred during the time roughly from 16,000 to 13,000 years ago.

[32] Johnston, "Pleistocene and Recent Deposits . . . . ," *op. cit.*, map.

[33] Antevs, "Retreat . . . . in Eastern Canada," *op. cit.*, pp. 67 and 95.

[34] Ragnar Lidén, "Den senkvartära strandförskjutningens förlopp och kronologi i Ångermanland," ("Manner and Chronology of the Late Quaternary Movement of the Shore Line in Ångermanland"), *Geol. Fören. i Stockholm Förhandl.*, Vol. LX (1938), pp. 397–404.

The subsequent transgression of some 50 feet, which led to the deposition of the upper marine clay and produced the Ottawa Sea beach standing at 240 feet at Ottawa and roughly 400 feet at Pembroke, is not readily explained. Under the assumption that the sea-level rose 240 feet during the age from 25,000 to 5,000 years ago, a rise of 50 feet, at the average rate of 12 feet in 1,000 years, would have required about 4,000 years. However, during the time roughly from 12,000 to 10,000 years ago there was an increase of the Labradoran and Patrician ice sheets (and probably of other ice sheets), as recorded by one very large and two or more small oscillations of the ice border in the region of Cochrane,[35] 250 miles north of Ottawa, suggesting that temporarily the rate of the rise of sea-level was at least subnormal. On the other hand, the increase of the ice load was possibly sufficient to press down the land somewhat at Ottawa in spite of the great strength and rigidity of the earth's crust.[36] Thus the Ottawa transgression was perhaps produced by a rise of the sea-level and a sinking of the land. If no land-sinking occurred, the quiescence may have lasted for more than 4,000 years. The Ottawa Sea may have reached its climax during the transition from the late Glacial to the Postglacial. The later events are not known at Ottawa.

LATE GLACIAL AND POSTGLACIAL UPWARPINGS OF THE GREAT LAKES

When the Fenelon Falls region began to rise after the three-outlet stage of Lake Algonquin had been in existence for thousands of years, the entire discharge was immediately diverted to the Port Huron outlet which, composed of clay, probably was cut below the Chicago outlet, which was held by a rock ledge. When the upwarping amounted to 40 feet at Cape Rich on the south side of the Georgian Bay, Lake Algonquin began to drain across the Algonquin highland to the Champlain Sea in the Ottawa Valley.[37] As lower outlets were uncovered during the retreat of the ice, the lake was gradually lowered to the threshold at North Bay, now standing at

[35] Antevs, "The Last Glaciation," *op. cit.*, pp. 149–50.

[36] B. Gutenberg, "Tilting Due to Glacial Melting," *Jour. Geol.*, Vol. XLI (1933), pp. 449–67; Daly, *op. cit.*, p. 121.

[37] Stanley, *op. cit.*, pp. 1678 and 1680.

690 feet altitude. The occurrence of an Algonquin beach at 1,178 feet 5 miles northeast of North Bay[38] suggests that the lowering totaled at least 488 feet. The drop of the water level to some 500 feet below the outlet at Port Huron gave rise to lake stages, of several thousand years' duration, the shores of which were drowned during the uplift of the North Bay outlet, except north of the pivot line from the North Bay outlet to Nipigon.[39] These lakes may be called the "Taylor Lakes" after Frank B. Taylor. During their existence an old river channel, which now is submerged 150–300 feet, was reopened between Lakes Michigan and Huron.[40]

The level of North Bay during the Ottawa Sea is not known; but subsequently the region rose probably a few hundred feet to within 102 feet of its modern stand and to the level of the Port Huron outlet, inaugurating the Nipissing Great Lakes. Just then equilibrium was restored, and quiescence prevailed probably for two to three thousand years, as suggested by the formation of the strong Nipissing beach, while both the North Bay and Port Huron outlets were in use.[41] The coincidence of this quiescence and of the Algonquin quiescence with two or three outlets was perhaps connected with the height of the lakes as determined by the capacity of the spillways. Thus the forces tending to raise the northern part of the lake basins, which would lead to the abandonment of the northern outlet, had to overcome an increase in the weight of the lakes caused by a rise and an expansion of their surface. A marine shoreline at 100 feet altitude near Montreal is perhaps a correlative of the Nipissing beach. Another possible correlative is the 20-foot Micmac terrace and sea cliff of the lower St. Lawrence,[42] whose remarkable strength may be due

[38] J. W. Goldthwait, "An Instrumental Survey of the Shore Lines of the Extinct Lakes Algonquin and Nipissing in Southwestern Ontario," *Geol. Surv. Canada Mem. 10* (1910), p. 33.

[39] Taylor, "The Pleistocene of Indiana . . . . ," *op. cit.*, pp. 457 and 461.

[40] Stanley, "The Submerged Valley through Mackinac Straits," *Jour. Geol.*, Vol. XLVI (1938), pp. 966–74.

[41] Taylor, "The Pleistocene of Indiana . . . . ," *op. cit.*, p. 449. (The so-called Nipissing beach north of the pivot line, at the extreme northeast corner of Lake Superior, began forming directly upon the final drainage of Lake Algonquin, which explains why it is stronger than the Nipissing beach proper [p. 457].)

[42] Goldthwait, "The Twenty-Foot Terrace and Sea Cliff of the Lower Saint Lawrence," *Amer. Jour. Sci.*, Vol. XXXII (1911), pp. 291–317.

TABLE 1

TENTATIVE CHRONOLOGY AND CORRELATIONS

Millenniums Ago*	†	Niagara Gorge	Outlets of Upper Basins	Upper Basins	Ontario Basin	Ottawa Lowland, Etc.
0–1–2–3 Late Postglacial	U ? ? ? U U U U	Upper Great gorge	Port Huron	Modern Great Lakes	Lake Ontario	
4–5–6 Middle Postglacial	Q Q Q Q Q	Ongiara gorge, and clearing	North Bay and Port Huron	Nipissing Great Lakes	Submerged lakes	
7–8 Early Postglacial	U U U U U	Niagara Glen gorge	North Bay	Taylor Lakes, submerged		
9–10–11–12–13 Younger late Glacial	Q Q Q S? S? S? Q Q Q U U U U U U U U					Ottawa Sea
14–15–16		Devil's Hole	Port Huron	Second Port Huron	Champlain Sea	
17–18–19–20–21 Middle late Glacial	Q Q Q Q Q Q Q Q Q Q Q U U U U U U U U	Old Narrow gorge	Fenelon Falls, Port Huron, and Chicago	Fenelon Falls–Port Huron–Chicago stage of Lake Algonquin	Lake Frontenac Lake Iroquois proper	Lake Vermont
22			Fenelon Falls	Fenelon Falls	Early Lake Iroquois	
22–23–24–25		First gorge	Port Huron	First Port Huron Lakes Lundy, Warren, Wayne, and Whittlesey		

* This chronology is based on varve (annual deposit of clay) counts and geological estimates, which are in part united by rough long-distance correlations of the temperature conditions, since it appears that the great, slow variations in the summer temperature influence the entire Northern, or Southern, hemisphere (Antevs, "Late-glacial Correlations and Ice Recession in Manitoba," *Geol. Surv. Canada Mem. 168* [1931], pp. 1–6; "Correlations of Late Quaternary Chronologies," *Report of XVI Internat. Geol. Cong.* [1936], pp. 213–16). The chronology is divided largely on the basis of variations in the summer temperature. According to this division, the Postglacial began when the temperature, rising from the last Glacial, had reached its modern level. The Middle Postglacial was a warmer (and drier) age than the present.

† U = Upwarping of the land. Q = Quiescence, stability. S = Sinking.

to transgression by rise of the sea-level above its modern stand during the warm Middle Postglacial.

Finally the North Bay region rose anew. As a consequence, the entire discharge was shifted to Port Huron, and the modern Lakes Huron, Michigan, and Superior were established. This occurred about 4,000 years ago—4,000 years being the estimated time needed by the modern Niagara River to erode the Upper Great gorge of the Niagara Gorge.[43] The same upwarping raised the outlet and the water body of the Ontario basin probably some 75 feet and formed modern Lake Ontario. The great recession of the Scarboro cliffs, the formation of Toronto Island, and the general strong development of the Ontario shoreline[44] are thus in part due to transgression and not exclusively to long time. Whether or not the mentioned upwarping was succeeded by quiescence is unknown, but North Bay is now rising at a rate of 60 cm. (23.6 inches) a century, or a foot in 50 years, while Chicago is stable.[45]

CORRELATIONS OF NIAGARA GORGE AND THE GREAT LAKES

The revised interpretation of the history of the Great Lakes obviously invalidates former correlations with the Niagara Gorge.[46] An attempt at a new correlation is presented in Table 1. The Devil's Hole gorge, the Niagara Glen gorge, and the Ongiara gorge, sections distinguished in the writer's former correlation, together form what Taylor called the Lower Great gorge. By "clearing" during the Nipissing Great Lakes age is understood the removal of the glacial debris which filled the Whirlpool, Eddy basin, Whirlpool Rapids gorge, and the Cantilever gorge, which were all cut prior to the Wisconsin glaciation.[47]

[43] W. H. Boyd, "A New Method of Determining the Rate of Recession of Niagara Falls," *Trans. Roy. Soc. Canada Sec. IV*, Vol. XXII (1928), pp. 1–12; Johnston, "The Age of the Upper Great Gorge of Niagara River," *Trans. Roy. Soc. Canada Sec. IV*, Vol. XXII (1928), pp. 13–29; see esp. p. 29.

[44] Coleman, "An Estimate of Postglacial and Interglacial Time in North America," *Compt. Rend. Congrès Géol. Internat. Canada, 1913* (Ottawa, 1914), pp. 435–49; "Glacial and Postglacial Lakes . . . . ," *op. cit.*, pp. 68 and 69.

[45] Gutenberg, *op. cit.*, p. 458.

[46] Taylor, "Niagara Falls and Gorge," *op. cit.*, p. 29; "New Facts on the Niagara Gorge," *Papers of Mich. Acad. Sci., Arts and Letters*, Vol. XII (1929), pp. 251–65; Antevs, "Late-glacial Correlations . . . . ," *op. cit.*, pp. 20–24.

[47] Johnston, "The Age of the Upper Great Gorge . . . . ," *op. cit.*, p. 26.

PHYSIOGRAPHIC SUBDIVISIONS OF THE SAN LUIS VALLEY, SOUTHERN COLORADO

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ABSTRACT

The San Luis Valley in south-central Colorado has heretofore been considered a single physiographic unit. In this paper are described five distinct geomorphologic subdivisions of the valley together with the fundamental geologic history which gave rise to the characteristics of each subdivision.

INTRODUCTION

The San Luis Valley, forming the upper end of the great valley of the Rio Grande, is one of the most impressive topographic features of southern Colorado. As originally outlined by Siebenthal,[1] it is a great lowland about 150 miles long and 50 miles in maximum width, flanked on the east by the linear Sangre de Cristo Range and on the west by the eastern portion of the more extensive San Juan Mountains. It is, in a sense, part of the chain of intermontane basins, or parks,[2] lying west of the Southern Rocky Mountain front ranges, but is unlike the others in having no southern mountain border. The meeting of the Sangre de Cristo Range and the San Juan Mountains about 135 miles north of the Colorado–New Mexico state boundary forms a natural northern limit to the valley. The southern limit is indistinct and was arbitrarily placed by Siebenthal about 15 miles south of the Colorado border (see index map, Fig. 1).

During the course of geological field work conducted during the summers of 1935, 1936, and part of 1937 in the southeastern part of the San Luis Valley, it became apparent that the valley is not a unit either geographically or geologically. To the south it actually merges with the great structural depression of the Rio Grande[3] in

[1] C. E. Siebenthal, "Geology and Water Resources of the San Luis Valley, Colorado," *U.S. Geol. Surv. Water-Supply Paper 240* (1910), p. 9.

[2] N. M. Fenneman, *Physiography of Western United States* (New York and London: McGraw-Hill Book Co., 1931), pp. 125–32.

[3] Kirk Bryan, "Outline of the Geology and Ground-Water Conditions of the Rio Grande Depression in Colorado and New Mexico" (mimeographed report, Rio Grande Joint Investigation, National Resources Committee, 1936).

New Mexico and within its own boundaries consists of five distinct subdivisions. Each of these smaller units, having had a somewhat different geologic history, now possesses distinctive geologic and topographic characteristics and is therefore considered as a distinct physiographic subprovince. In the following pages is presented a description of these subprovinces and a brief account of their

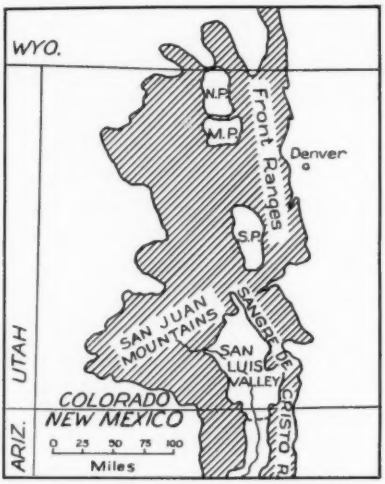

FIG. 1.—Index map showing the position of the San Luis Valley in the Southern Rocky Mountains (shaded) and its relation to the intermontane parks: *N.P.*, North Park; *M.P.*, Middle Park; *S.P.*, South Park. Boundaries from Fenneman, slightly modified.

geologic backgrounds. The subdivisions described are named and located as follows: the Alamosa Basin, the main northern part of the San Luis Valley; the San Luis Hills, forming most of the southeastern border of the basin; the Taos Plateau, extending southward from the San Luis Hills; the Costilla Plains, constituting an alluvium-mantled strip along the east side of the Taos Plateau; and, finally, the Culebra Re-entrant, that part of the San Luis Valley lying east of the previously named areas (Fig. 2). Many of the geomorphologic interpretations herein presented are the result of fairly detailed structural and stratigraphic work. The writer intends to

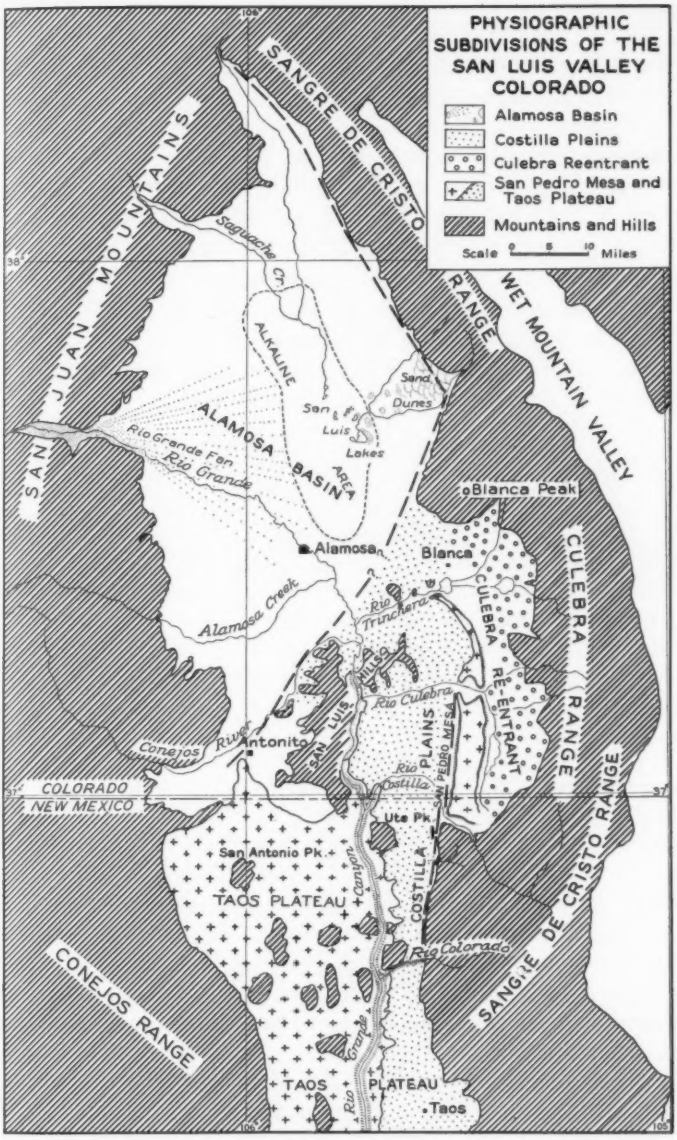

FIG. 2.—Physiographic subdivisions. Approximate location of the principal late faults shown by heavy, dashed lines. Two recent scarplets east and southeast of Ute Peak indicated by short, heavy lines.

discuss those studies directly in later papers. They will be little more than mentioned here.

Dominating the San Luis Valley region is the high and rugged Sangre de Cristo Range along the east side. The range consists of two parts: the northern part extending in a general southeasterly direction for 50 miles and the southern part pursuing a southerly or slightly southwesterly course for an additional 18 miles. From north to south the range crest gradually increases in elevation, culminating east of Alamosa in a mass of rugged peaks separated from each other by cirques and glaciated valleys. Highest of these is Blanca Peak, 14,390 feet above sea-level. The entire group of peaks may be designated by the Spanish term, "Sierra Blanca" (Fig. 3).

FIG. 3.—Sierra Blanca from the southwest showing the glaciated valleys and the bordering alluvial fans. The level plain in the fore- and middle-ground is that in which the Alamosa Basin merges with the Costilla Plains and the Culebra Re-entrant.

From Sierra Blanca northward to Poncha Pass, at the junction of the Sangre de Cristo Range with the San Juans, the range possesses a bold, linear west front undoubtedly indicative of relatively recent uplift along normal faults. Not even the accumulation of considerable debris in the form of alluvial fans can modify the sharp topographic break between the abrupt mountain slope and the relatively flat valley floor. South of Sierra Blanca, however, the main part of the range is sharply offset to the east about 15 miles, and the steep and abrupt character of the mountain front is conspicuously absent. The front here is subdued and highly irregular, consisting of long mountain spurs which merge westward with an extensive belt of foothills.

The offset portion of the range is about 40 miles long and extends in a southerly direction to a little beyond the Colorado–New Mexico border. Near that boundary the Sangre de Cristo Range is offset

westward and exhibits the abrupt, straight, and steep western front like that of the range farther north and similarly indicative of considerable displacement on one or more faults. Thus immediately west of the eastwardly offset part of the range is a sort of indentation, or re-entrant. The offset part of the range bears the local name, "Culebra Range," and the re-entrant is herein named the "Culebra Re-entrant." It is geologically the most significant of the physiographic subdivisions to be subsequently described, as in it may be seen rock bodies and structural relationships which do not seem to be visible elsewhere in the area. The subdivisions of the valley will be defined and described in the order in which they were previously listed.

ALAMOSA BASIN

Siebenthal's work in the San Luis Valley was largely restricted to the main northern part of the area lying north of the San Luis Hills. He appears to have considered the other portions of the valley peripheral and of little importance. From the standpoint of water supply this view is justified, but, in a physiographic consideration of the San Luis Valley as a whole, the area emphasized by Siebenthal plays a more subordinate role. It is therefore proposed to make that area a physiographic subdivision of the valley as herein defined and to name it the "Alamosa Basin" after the principal town, Alamosa.

The Alamosa Basin is a roughly triangular-shaped area bordered on the west by the San Juan Mountains and on the northeast by the Sangre de Cristo Range. The southeast side is incompletely marked, partly by the Sangre de Cristo Range and partly by the San Luis Hills. Between these hills and the southwest end of the Sangre de Cristo Range is a low, nearly flat area (Fig. 3), across which the Alamosa Basin merges imperceptibly with two of the other physiographic subdivisions of the area—the Costilla Plains and the Culebra Re-entrant.

The principal characteristics of the Alamosa Basin are its nearly featureless floor, the recency of the deposits in the basin, and the remarkable course of the Rio Grande directly across the area from the west to the southeast sides. Actually, most of the valley floor slopes inward on all sides toward the lowest portion near the eastern

margin. On the east side are numerous alluvial fans whose heads lie at the mouths of short, precipitous canyons in the Sangre de Cristo Range. Streams entering the basin from the west are much longer, have more extensive drainage basins in the San Juan Mountains, and have developed much broader and more gently sloping alluvial fans. At a few miles north of Alamosa an observer appears to be surrounded by a level plain. This seeming plain is actually the surface of the Rio Grande fan—the slope so gentle as to be perceptible only on a topographic map. Thus, underlain directly by modern alluvial fan deposits, the Alamosa Basin is essentially an area of deposition. Only at the north where San Luis Creek flows out of Poncha Pass and at the south where the Rio Grande leaves the area through the San Luis Hills is there any dissection of the alluvial surface.

The visible deposits in the Alamosa Basin constitute the upper layers of the Alamosa formation which was originally named by Siebenthal[4] and held by him to be of Pleistocene age. Beds of the formation are known only from well-logs and from a few scanty exposures near the south end of the basin where the Rio Grande has cut a shallow trench. The sediments appear to be nearly flat-lying but possess sufficient centripetal inclination to make a good artesian basin, from which most of the water supply of the northern part of the San Luis Valley is derived. As indicated by well-logs, the deposits are at least 4,000 feet thick below the lowest part of the basin. Deposition was apparently accompanied by downward tilting of the valley block along great faults bordering the Sangre de Cristo Range and the northwest side of the San Luis Hills (Fig. 2). Evidence for the faulting is mainly topographic and was recognized by Siebenthal,[5] Atwood and Mather,[6] Cross and Larsen,[7] and others.

Partly because of the location of the faults, and partly because of the huge size of the western alluvial fans, the lowest part of the basin is near the eastern margin. Here are shallow lakes which have

[4] *Op. cit.*, p. 40.

[5] *Ibid.*, pp. 38–39 and 51.

[6] W. W. Atwood and K. F. Mather, "Physiography and Quaternary Geology of the San Juan Mountains, Colorado," *U.S. Geol. Surv. Prof. Paper 166* (1932), pp. 23, 25, and 99.

[7] Whitman Cross and E. S. Larsen, Jr., "A Brief Review of the Geology of the San Juan Region of Southwestern Colorado," *U.S. Geol. Surv. Bull. 843* (1935), p. 113.

become somewhat alkaline through evaporation (Fig. 2). In fact the central part of the basin is known as the "Alkaline Area" because of the alkalinity of the water. Wind-blown sand has accumulated around the lakes and, in times of relative aridity, seems to be subject to deflation by the prevailing southwesterly winds. As a result, considerable sand has been piled up at the mountain front near by.

FIG. 4.—Sand dunes of the Great Sand Dunes National Monument

This body of sand has recently been set aside by the government as the Great Sand Dunes National Monument (Fig. 4). The writer, jointly with Professor H. T. U. Smith of the University of Kansas, has begun a more detailed study of these dunes.

SAN LUIS HILLS

The Alamosa Basin is nearly completely separated from the rest of the San Luis Valley to the south by the San Luis Hills. They constitute a fairly rugged mass of hills and tilted mesas 500–1,000 feet high, extending from near the town of Antonito, Colorado, across the Rio Grande to near the town of Blanca, Colorado. Beginning at a broad pass near Antonito they rise to their full height above the surrounding plain. Toward the northeast they gradually

decrease in altitude to pass beneath alluvium a few miles south of Blanca. The continuity of the hills as a single mass is broken only by the canyon of the Rio Grande, beginning a few miles south of Alamosa.

The San Luis Hills were apparently carved out of dominantly andesitic volcanic rocks correlated with portions of the mid-Tertiary volcanic series of the San Juan Mountains after the development of the San Juan peneplain of Atwood and Mather. The dissection of the hills resulted from postpeneplain uplifts on normal faults situated along their northwest flank.[8] The present writer believes that the faults are probably the southward continuation of those bordering the west flank of the Sangre de Cristo Range. Recurrent movements accompanied the development of the Alamosa Basin and at the same time served to maintain the hills as a prominent topographic feature.

After dissection had proceeded to the development of a mature topography, the San Luis Hills were surrounded by basalt flows belonging to the Hinsdale series.[9] The basalt extended over large areas in the San Juan Mountains and for many miles down the Rio Grande Valley to the south. At one time the hills rose fairly abruptly from the surrounding flood of lava—a feature at present preserved only along their southern side. On the northwest side of the hills the lava is buried by beds of the Alamosa formation, which rests with fault contact against the older mid-Tertiary volcanics.

Because of the island-like relationship of the San Luis Hills to the lavas, they might be considered as part of the Taos Plateau—the next subdivision to be discussed. There hills of older rocks rise above the Hinsdale flows as do the San Luis Hills, and many of them appear to be composed of volcanics of the same general series. However, the San Luis Hills are not isolated from each other as are those farther south but form a nearly continuous group both topographically and apparently also lithologically. In addition the Hinsdale has been downfaulted along the northwest side of the hills. They are therefore considered as a distinct physiographic subdivision but

[8] Atwood and Mather, *op. cit.*, p. 23.

[9] Cross and Larsen, *op. cit.*, pp. 94–100; Atwood and Mather, *op. cit.*, p. 21.

are not, however, distinguished from other mountains and hills in Figure 2.

TAOS PLATEAU

The Taos Plateau is an extensive plateau-like area which lies south of the San Luis Hills and extends about 60 miles southward into New Mexico. Its distinguishing features are a broadly undulating character, the presence of rounded hills rising above the general level, and the deep entrenchment of the Rio Grande and other streams. The name was suggested by Kirk Bryan from the clear development of its broad, undulating character near and west of Taos, New Mexico. The plateau is largely underlain by basalt belonging to the Hinsdale series. The hills rising above the rolling surface of the lava are composed of older Tertiary and pre-Cambrian rocks, against which, in several places, the basalt may be seen to rest unconformably.[10]

The Taos Plateau is trenched near its eastern side by the Rio Grande Canyon which, only about 50 feet deep at the mouth of the Rio Culebra in Colorado, becomes progressively deeper to the south. At the mouth of the Rio Costilla the river is approximately 200 feet below the surface of the lava (Fig. 5). At its southern end, near Embudo, New Mexico, the canyon is about 1,000 feet deep. There the plateau-like character of the physiographic subdivision is most apparent.

On the west side of the Rio Grande Canyon the basalt essentially forms the surface of the plateau except for the projecting hills of older rocks. As the basalt is slightly deformed, so also is the plateau surface. The lava is known to be broken by normal faults near the western margin of the northern portion of the plateau;[11] and in the vicinity of Dunn's Bridge, across the Rio Grande northwest of Taos, the basalt is believed by Kirk Bryan[12] to be slightly deformed. Because of the regional eastward tilt of the lava the surface of the plateau is somewhat higher west of the Rio Grande than east of it.

East of the Rio Grande the plateau exhibits different features in

[10] Bryan, "Geology of the Rio Grande Canyon," in preliminary report of the New Mexico state engineer (1928–30).

[11] A. P. Butler, Jr., Doctor's thesis (Harvard University) in preparation.

[12] Personal communication.

that the basalt has been buried beneath a considerable accumulation of alluvium derived from the Sangre de Cristo Range. However, from the Rio Colorado southward the area is deeply trenched by large canyons made by the Rio Colorado itself, by the Rio Hondo, and by other streams issuing from the Sangre de Cristo Range. This dissection emphasizes the plateau-like character of the region sufficiently to warrant its inclusion as part of the Taos Plateau in spite of the alluvial cover on the basalt.

FIG. 5.—Rio Grande Canyon about ½ mile south of mouth of Rio Costilla. Canyon walls, developed in Hinsdale basalt underlying the Taos Plateau, exhibit evidence of two stages of erosion.

The upper layers of the alluvium, as exposed in canyon walls near the Rio Grande west of Taos, clearly overlie the basalt. Farther up the tributary streams, however, the lava is either missing or not exposed, and the precise relationships of the alluvium there are unknown. It may contain parts of more than one formation.

North of the Rio Colorado the surface of the alluvium is very little dissected and the area is considered as a separate physiographic subdivision.

COSTILLA PLAINS

The area flanking the Sangre de Cristo Range north of the Rio Colorado, lying east of the Rio Grande, and extending northward to

a little beyond the Rio Trinchera in Colorado, is called the Costilla Plains. The term is derived from the town of Costilla, New Mexico, situated where the Rio Costilla crosses the Colorado–New Mexico boundary near the mountain front. In this subdivision the alluvial deposits blanketing the eastward extension of the Hinsdale basalt are practically undissected except at the very mouths of the streams tributary to the Rio Grande. But for the slight westward slope of the surface of the alluvial fans heading in the canyons of the Sangre de Cristo Range, the area is nearly featureless. North of the Colorado–New Mexico boundary it is nearly a level surface.

The Costilla Plains are only about 4 miles wide near their southern end but are considerably wider in Colorado where they reach a maximum of about 15 miles a few miles north of the New Mexico border. Their western limit is strictly the edge of the alluvium, but that is very near the Rio Grande Canyon, so that natural barrier is made the western border of the Costilla Plains in Figure 2. On the east the plains abut against the southern portion of the Sangre de Cristo Range and against the San Pedro Mesa (Fig. 2). Corresponding as it does with the western front of the mountain range, the east border of the plains is nearly a straight line and probably marks or closely parallels the trace of the normal fault along the edge of the mountains. The alinement of the west side of the San Pedro Mesa with this mountain scarp to the south suggests that the concealed fault continues to or beyond the north end of the mesa.

More definite evidence of the existence of the fault is furnished by two small scarplets in the alluvium. One of these is about 70 feet high and crosses the upper portion of the alluvial fan of La Jara Creek approximately at the mountain front. La Jara Creek is the small, unnamed stream southeast of Ute Peak, shown in Figure 2. The other scarplet is somewhat longer and lower, being about 30 feet high and extending for about a mile across the lower portion of the fan of Cedros Creek—the next stream north of La Jara Creek. It is not shown in Figure 2, although the location of the scarplets is indicated. The longer scarplet approximately parallels the mountain front and can be readily seen a few hundred yards east of the road to Taos and 3 or 4 miles south of the Colorado state line (Fig. 6).

Whereas most of the surface of the Costilla Plains is untrenched

by streams, Rio Trinchera, Rio Culebra, and Rio Costilla in Colorado have formed floodplains about 100 feet below the general surface. That surface therefore constitutes a terrace. There are also remnants of a lower terrace about 55 feet above present stream grades. The generally fine, sandy, unconsolidated deposits visible in a few places in the sides of the terraces are tentatively correlated with the Alamosa formation of Siebenthal in the Alamosa Basin.

FIG. 6.—Slightly dissected Recent fault scarp across alluvial fan and nearly parallel to the west front of the Sangre de Cristo Range near Costilla, New Mexico. View looking northeast.

Topographically the Costilla Plains are very similar to the Alamosa Basin but are believed to be different in that at least the Colorado portion of their surface is mainly erosional, whereas the surface of the Alamosa Basin is largely depositional. Additional cause for separating the Costilla Plains from the Alamosa Basin lies in the fact that they are almost entirely cut off from the basin by the San Luis Hills. However, the Costilla Plains, the Culebra Re-entrant, and the Alamosa Basin actually merge together in the small area west of the town of Blanca. For that reason physiographic boundaries are not drawn in that area on Figure 2.

CULEBRA RE-ENTRANT

The Culebra Re-entrant differs markedly from the other physiographic subdivisions of the San Luis Valley. The topography is considerably diversified and exhibits a mature stage of dissection. In contrast to the San Luis Hills, which also possess a mature topography, the hills of the re-entrant have not been surrounded or buried by basalt flows.

Although the Culebra Re-entrant is somewhat isolated and appears to be insignificant in comparison with the topographically conspicuous and economically important Alamosa Basin, it is geologically very important. In the re-entrant are exposed formations of which only one, or at most two, crop out in any one of the other subdivisions. The formations of the other areas, however, do occur in the re-entrant with the exception of the Alamosa formation, which is represented only by local alluvium. Consequently, only in the re-entrant can the structural relations of all the rock bodies of the region be satisfactorily determined.

The pre-Cambrian rocks of the Culebra Range have been uplifted on a series of normal faults separating the range from the area to the west, which is underlain by deformed Tertiary volcanic rocks and fluvial deposits. The topographic border of the mountains follows the fault boundary fairly closely, but in the southern part is very complicated. It is generalized in Figure 2. The term "Culebra Re-entrant" is restricted to the area lying west of the faults. In that area three formations are exposed: (1) the early Tertiary Vallejo formation,[13] hitherto undescribed; (2) an overlying body of flows and tuffs; and (3) a still younger, thick body of alluvial-fan sand and gravel interbedded with the basalt that elsewhere surrounds the San Luis Hills. These rocks and their structural relationships are significant here mainly because as a body the rocks were deformed and subject to erosion considerably before the fault movements along the east side of the Alamosa Basin and the Costilla Plains had ceased. Indeed, as previously indicated, movement on these faults has been very recent, and the deformation may be thought of as continuing into the present. Because movement on the faults of the

[13] J. E. Upson, "The Vallejo Formation—New Early Tertiary Red-Beds in Southern Colorado" (in preparation).

Culebra Re-entrant ceased relatively early, that area became one of dissection, while deposition continued in the Alamosa Basin and Costilla Plains. The comparatively great length of time since the erosion of the re-entrant rocks began is the reason for the mature topography of that area in contrast to the nearly featureless Alamosa Basin and Costilla Plains and the youthfully trenched Taos Plateau. The mature topography of the San Luis Hills may be considered as a fossil topography preserved by the flood of younger basalt. When the dissection of the Taos Plateau began is not known and is an inviting field for further study.

The Culebra Re-entrant is herein considered as a single physiographic subdivision because it seems to have had the same geologic history in all parts. It is, however, a topographically diversified area. Differences are largely the result of the details of the late Tertiary normal faulting acting on the volcanics and fluvial deposits. The major trend of the normal faults is north-south and movement on the faults was such as to divide the area into three parts: (1) a relatively elevated belt of foothills near the mountains; (2) a central graben, subsequently eroded so as to preserve the original topographic depression; and (3) a prominent horst—the San Pedro Mesa.

The eastern border of the belt of foothills is not conspicuous, as those hills merge with the long spurs and ridges cut out of the pre-Cambrian rocks of the Culebra Range proper without prominent topographic break. The latest movement on the faults separating the range from the foothills was probably not younger than early Pleistocene, and subsequently erosion has greatly modified the fault scarps that must once have existed. As a result the western front of the Culebra Range is noticeably less abrupt and imposing than the corresponding fronts of the Sangre de Cristo Range, both north and south of the re-entrant. The range crest itself, however, is in some respects more rugged than those to the north and south.

The foothills are widest in the northern part of the re-entrant where their western edge is 8–10 miles from the mountain border on the east. In the southern part of the area their width decreases from about 6 miles near the Rio Culebra to nearly zero at the Rio Costilla. The main feature of the foothill belt is the preservation of remnants of at least four erosion surfaces cut across the deformed

Tertiary rocks. In the north very few residuals rise above the highest erosion remnants, which lie at 450–500 feet above the present stream grades.

Remnants of apparently the same erosion surfaces are recognizable, but less extensively preserved, in the southern part of the re-entrant. There numerous foothill summits, rising as much as 10,000 feet above sea-level, are residual above the highest erosion remnants.

Extensive development of comparable erosion surfaces is not known in the other subdivisions of the San Luis Valley. In a few localities, however, there are small remnants of terraces, possibly erosional, which may be the correlatives of the surfaces in the re-entrant. Accurate information regarding such possible correlatives is not now at hand.

As a sort of outlier of the northern part of the foothill belt in the Culebra Re-entrant are two prominent basalt-capped mesas called the Garland Mesas, situated immediately southeast of the town of Fort Garland, 5 miles east of Blanca. The southern of the two mesas rises over 500 feet above the surrounding lowlands. The rocks of these mesas have been uplifted on a normal fault of considerable magnitude, first recognized by Atwood and Mather.[14]

Extending northward from the southwest corner of the re-entrant is the similarly basalt-capped mesa, the San Pedro Mesa. It is 1,000 feet high at the south end and forms a prominent table-land, particularly when viewed from the west, extending northward for about 15 miles from the Rio Costilla to the Rio Culebra. Near the Rio Culebra the mesa decreases in elevation and the capping basalt assumes a gentle westerly dip. The basalt outcrop crosses the Rio Culebra and, maintaining the westerly dip, forms a narrow, cuesta-like ridge, extending for about 10 miles north of Rio Culebra and presenting a low but fairly steep escarpment toward the east. This ridge is the northward continuation of the San Pedro Mesa and is called the San Pedro Cuesta in this paper. It dies out a short distance north of the Rio Trinchera.

The San Pedro Mesa is apparently bordered by faults on both the east and the west sides. As the upthrow sides of each of these faults is on the mesa side, it is structurally a horst. The fault on the east side continues north along the foot of the San Pedro Cuesta.

[14] *Op. cit.*, p. 100.

There are also a few small normal faults along the western margin of the foothills with downthrow on the west. The region between the San Pedro Mesa and the foothills is therefore structurally a graben. The fault movements are believed to have carried the basalt of the mesas down so as to render the graben easily eroded. At present the feature is topographically fairly conspicuous and was designated by Stephenson,[15] the "Culebra Park." The writer sees no reason to change this name. The Culebra Park extends in a northerly direction from near the Rio Costilla, where it is very shallow but where its floor is about 500 feet above the level of that stream. This relationship suggests that at some previous time the upper Rio Costilla flowed northward through the Culebra Park but was diverted by the present Costilla cutting headward across the south end of the San Pedro Mesa from the west. North of Rito Seco the park is broader and deeper, assumes a northwesterly trend, and extends past the town of Blanca to merge with the surface of the Alamosa Basin in the northwest part of the re-entrant.

North of the Rio Costilla the surface of the Culebra Park is about at the level of the present stream grades except in the region between Rito Seco and Rio Trinchera. There the floor of the park is 20–100 feet above the level of the streams, and on the flanks of the divide thus formed low terraces may once have been developed. They are at present so mantled by later alluvium as to be unrecognizable.

ACKNOWLEDGMENTS.—Acknowledgments are due principally to Professor Kirk Bryan of Harvard University, who, long familiar with the region, first suggested the possibility of subdividing the San Luis Valley. Professor Bryan criticized the manuscript many times in its formative stages, aided much in the drafting of Figure 2, and is the originator of some of the subdivision names. Mr. Edward Schmitz also contributed to the preparation of Figure 2. The writer is indebted to many inhabitants of the Culebra Re-entrant for assistance and hospitality during the field work there—notably, Mr. and Mrs. S. H. Shannon and family of the Trinchera Ranch, Fort Garland. During the field work the writer was ably assisted at different times by H. G. Peacock, A. P. Butler, Jr., W. S. Brandhorst, R. W. Day, and the writer's brother, D. R. Upson.

[15] J. J. Stephenson, "Report on the Geology of a Portion of Colorado Examined in 1873," *U.S. Geog. Surv. West of the 100th Meridian*, Vol. III (Supplement Geology, 1881), p. 363.

GLACIAL GEOLOGY OF THE SUNLIGHT AREA PARK COUNTY, WYOMING[1]

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ABSTRACT

Sunlight Creek flows northeastward into Clarks Fork River in the northern Absaroka Mountains of northwest Wyoming. Two periods of glaciation are here recognized from the degree of erosion of the moraines. An early ice sheet from the Beartooth Plateau to the north invaded Clarks Fork Valley; but, being unable to move eastward through the Clarks Fork Canyon to the plains as rapidly as ice filled the valley, it moved south against the Absaroka Range, filling the lower ends of valleys in that range and even overflowing the lower passes in the Clarks Fork–Sunlight divide. This ice and its moraine dammed the Sunlight Valley and formed a lake, the aggradation of which has floored the Sunlight basin. The later glaciation was of the normal valley type with a glacier extending part way down Sunlight Creek Valley. A few small glaciers still exist today.

INTRODUCTION

The Sunlight area is located in the northern Absaroka Mountains in the Shoshone National Forest, Park County, Wyoming, just south of the Beartooth Plateau. Most of the features herein described are found along Sunlight Creek and its tributaries, a northeastward-flowing stream which joins the Clarks Fork River at the southeast corner of the Beartooth Plateau. The western part of the area was first mapped by Hague in 1893 and 1897.[2] Dake[3] made note of some of the glacial features of the region in 1919. The field work for the present study was done during the summers of 1934 and 1935 and represents one of the co-operative studies of the Yellowstone-Bighorn Research Association.

GENERAL GEOLOGIC SETTING

The southern flank of the Beartooth uplift, with Paleozoic formations assuming very steep southerly dips in places with local fault-

[1] Presented at the Fiftieth Annual Meeting of the Geological Society of America in Washington, D.C., December, 1937. W. H. Parsons, "Glaciation in the Sunlight Valley, Wyoming" (abst.), *Geol. Soc. Amer. Proc. 1937* (1938).

[2] Arnold Hague, Absaroka Folio, No. 52, *U.S. Geol. Surv. Atlas* (1899).

[3] C. L. Dake, "Glacial Features on the South Side of Beartooth Plateau, Wyoming," *Jour. Geol.*, Vol. XXVII (1919), pp. 128–31.

ing,[4] is just north of the Clarks Fork River and on Bald Ridge. In consequence of this uplift pre-Cambrian granite outcrops only 3 or 4 miles north of the Clarks Fork River at elevations up to 10,600 feet, or some 4,000 feet above the Sunlight Valley.

The Clarks Fork River, flowing in an easterly direction and lying just a few miles north of the Sunlight area, has cut into pre-Cam-

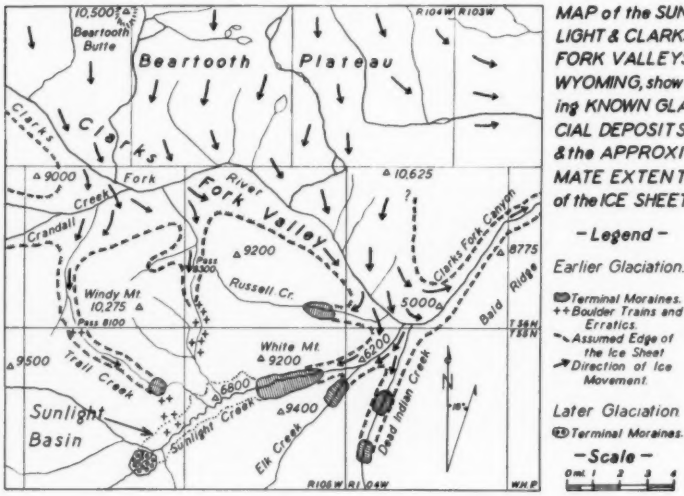

FIG. 1

brian granite, and Sunlight Creek itself flows in granite for about 1½ miles before joining the Clarks Fork. Along this river and south of it, with the exception of Bald Ridge, all remnants of Cambrian formations overlying the pre-Cambrian are essentially flat-lying.

The rocks of the Sunlight area consist essentially of flat-lying Paleozoic (Cambrian through Mississippian) limestones overlain by Tertiary andesitic pyroclastics and lava flows. These are cut by Tertiary plugs, laccoliths, small stocks, and numerous radiating dikes.[5]

[4] Hague, *op. cit.;* R. V. Hughes, "The Geology of the Beartooth Mountain Front in Park County, Wyoming," *Proc. Nat. Acad. Sci.*, Vol. XIX (1933), pp. 239–53.

[5] Parsons, "Volcanic Centers of the Sunlight Area, Park County, Wyoming," *Jour. Geol.*, Vol. XLVII (1939), pp. 1–26.

This area is characterized by rolling hills and some rocky spurs with a relief near Sunlight Creek of 1,000–2,000 feet, increasing to 3,000 feet a few miles back from the main valley. The Sunlight Valley in its eastern part is nearly a mile wide, very flat, and called Sunlight basin. Here the creek meanders very extensively (Fig. 2) over the wide swampy valley. Several of the tributary valleys are also wide and flat but do not have permanent streams.

FIG. 2.—Sunlight basin. View upstream showing the meandering character of Sunlight Creek.

GLACIATION

Two stages of Pleistocene glaciation have been recognized in the Sunlight area. The earlier of these may possibly be correlated with the early Wisconsin. The younger glaciation, in late Wisconsin time, was responsible for the final sculpturing of the high mountain topography but left few depositional features in the Sunlight area.

THE EARLIER GLACIATION

In the lower Sunlight Valley, below the Sunlight basin and beside White Mountain, is a large moraine characterized by a gentle swell and swale topography through which the creek has cut along the north side of the valley. This moraine, which is about 2 miles long and as wide as the valley (over ½ mile), rises gently in an even sweep about 50 feet above the level of Sunlight basin. The rolling

topography of the moraine, cut by the V-shaped valley of Sunlight Creek, makes a striking contrast to the flat-bottomed Sunlight basin immediately to the southwest, where the creek meanders extensively (Fig. 2) over wide gravel bars and through swampy fields.

The moraine is composed of a heterogeneous boulder clay with various-sized pebbles and boulders in a light-colored matrix without any sort of bedding. The pebbles average from an inch in diameter to 6 inches or a foot, though a few large boulders are as much as 6 or 10 feet in diameter. The larger boulders are well rounded, while the smaller ones are more subangular, and many show good glacial striations. The rock composition of the pebbles between 1 and 6 inches in diameter is given in Table 1. Of the boulders over 2 feet in diameter, 90 per cent are composed of granite.

TABLE 1

	Percentage
Pre-Cambrian granite	10
Paleozoic limestone	72
Sandstone and quartzite	10
Volcanic rocks	8
Total	100

In the valley of Trail Creek, a southeast-flowing tributary of Sunlight basin, occurs another and smaller moraine belonging to the same glacial episode. The moraine is nearly 2 miles from Sunlight Creek and is about $\frac{1}{2}$ mile long and equally wide, rising 200 feet above the valley floor against the side hills. It is cut in two by a postglacial valley 300 feet wide. The composition of this drift differs from that of the Sunlight moraine. Pebble counts give the percentages shown in Table 2. Here also most of the boulders over 2 feet in diameter are of granite.

TABLE 2

	Percentage
Pre-Cambrian granite	4
Paleozoic limestone	25
Sandstone and quartzite	4
Volcanic rocks	67
Total	100

Other similar but smaller moraines occur on Russell, Elk, and Dead Indian creeks, east- and north-flowing tributaries of the Clarks Fork Valley. These occurrences have already been noted by Dake.[6]

In the valley of Painter Creek, another south-flowing tributary of Sunlight basin lying to the east of Trail Creek, are many granite boulder trains. These boulders are 4–5 feet in diameter. Similar granite erratics are scattered over the low foothills on the south side of Trail Creek Valley.

The composition of all this glacial material indicates that it could not have been brought from the Absaroka Mountains to the west. Some Paleozoic limestone outcrops do occur in the Sunlight, Russell, Dead Indian, and Elk Creek drainage basins, but no pre-Cambrian is exposed there. If the Sunlight moraine, for example, had been deposited by a glacier coming down Sunlight Valley, it should contain about 90 per cent of volcanic rock instead of less than 10 per cent. Therefore, these moraines must have been deposited by glaciers originating outside the Absaroka drainage system.

The only pre-Cambrian–Paleozoic outcrop areas from which material of the observed composition could have been derived are to the north in the Clarks Fork Valley and the Beartooth Mountains. Therefore, glacier tongues must have moved up Sunlight, Dead Indian, and Russell creeks from the Clarks Fork Valley.

From the much glaciated Beartooth Plateau to the north a large ice sheet probably spread down into the wide Clarks Fork Valley, filling it with a great depth of ice. This great mass of ice from the north abutted against the Absarokas to the south; but, further advance in that direction being possible only up a few small tributary valleys, the main ice tongue turned eastward and moved down the Clarks Fork Valley. The latter valley is very wide where Russell, Sunlight, and Dead Indian creeks join it. Just east of the mouth of Dead Indian Creek, however, the valley abruptly narrows where the river has cut a major canyon through the upturned beds of the Beartooth Mountain front. This abrupt narrowing acted as a partial dam, and only a narrow ice tongue was able to move north-

[6] *Op. cit.*

eastward through the Clarks Fork Canyon onto the edge of the plains. The rest of the ice continued south and turned southwest into Absaroka valleys, such as the Sunlight. The assumed extent of this ice sheet and its approximate directions of movement are represented in Figure 1.

The moraine in Trail Creek Valley, however, was not formed by a glacier moving up from Sunlight basin. Such action would have required the filling of Sunlight basin and its tributaries with ice, and we should find moraines in other tributary valleys similar to the one in Trail Creek Valley. No such moraines exist.

From the headwaters of Trail Creek a relatively low pass (elevation 8,100 feet) leads to the headwaters of a north-flowing stream of the Clarks Fork system. On this divide, J. T. Rouse[7] observed a large number of gneiss and granite boulders, some up to 4 feet in diameter, together with fragments of Lower Paleozoic rocks about 1,000 feet above the volcanic-Paleozoic contact. Evidence shows, he says, that these boulders have not weathered out of the volcanic breccias. They must have been transported by ice. We conclude, therefore, that a tongue of the Clarks Fork glacier pushed south up over this pass and 4 miles down Trail Creek, there depositing the moraine this valley possesses.

The boulder trains in Painter Creek Valley also indicate that a glacial arm from the Clarks Fork pushed over a low divide (elevation 8,300 feet) but did not descend far beyond the divide. Only outwash was carried into the valley of Painter Creek. The Painter and Trail Creek passes are the only low ones between the Sunlight and Clarks Fork valleys.

The surfaces of all these moraines are fairly smooth and very gently rolling (Fig. 3.). This character indicates a considerable amount of erosion of the moraines to reduce them from the typical knob-and-kettle topography of more recent moraines. Late Wisconsin moraines generally show a very hummocky surface but little altered from the original rough surface of deposition. Because of the lack of hummocky topography, the Sunlight moraines are probably older that late Wisconsin.

[7] Personal communication.

THE LATER GLACIATION

In a narrow part of the Sunlight Valley above Sunlight basin is a moraine with a very uneven, hummocky surface having a relief of over 100 feet. It is about ½ mile in length and width and is cut through the middle by Sunlight Creek. Its gravels are composed mainly of volcanic rocks with a few fragments of Paleozoic limestone but no granite. It clearly was deposited by a valley glacier coming down Sunlight Valley from the crest of the Absaroka Range.

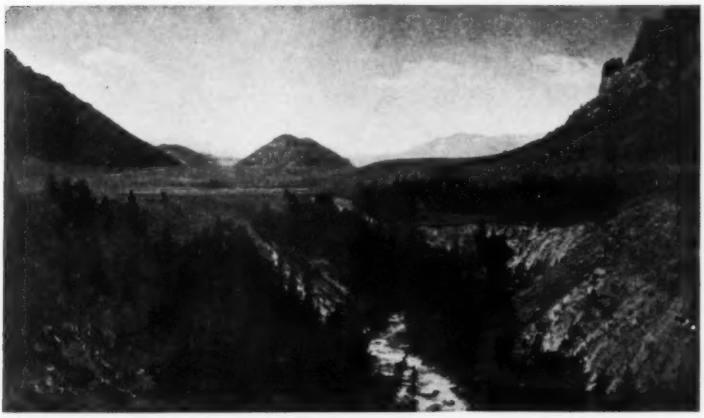

FIG. 3.—The older moraine (below Sunlight basin). Center background shows high part with wide profile valley; foreground shows lower part with narrow steep-sided valley.

The moraine is considered late Wisconsin in age because of its uneven and only slightly eroded topography and because it rests on the alluvial fill of the Sunlight basin. In this it is in marked contrast to the moraine in the lower Sunlight Valley, which has a fairly even topography and is the dam which caused the deposition of the alluvial fill.

A few small glaciers exist today in the Sunlight area. Of these, the Sunlight glacier is by far the largest, being a little over a mile long and averaging more than ½ mile wide. It lies at the foot of Sunlight Peak in a large composite cirque basin at the head of Sulphur Creek, between 10,000 and 11,000 feet elevation. Its lower end is well

covered by moraine. The other three or four glaciers in the area are tiny cliff glaciers plastered on the southwest walls of northeast-facing cirques and average not quite 2,000 feet in maximum dimension.

AGGRADATION OF SUNLIGHT BASIN

The ice tongue coming from the Clarks Fork Valley dammed the Sunlight Valley and formed a lake in the Sunlight basin. Sedimentation started in this lake as soon as it was formed. The tributary streams were heavily laden with outwash from glaciers in the higher Absarokas, and the glacier tongue from the Clarks Fork also contributed to the rapid sedimentation.

FIG. 4.—Flat alluvial fill of stream tributary to Sunlight basin

During the stay of the ice, lake-level must have been changeable, as the lake discharged over and around the edge of the ice tongue. No shoreline terrace remnants have been found to indicate any long-maintained water-level. When the ice retreated, the level of the lake sank to the lowest point in the morainal dam. Probably a stream soon cut into the moraine, lowering lake-level still more.

The rivers continued aggrading after the lake was filled with sediment, especially in the upper Sunlight Valley and its tributaries. Alluvial fill was gradually extended up the side streams. The Sunlight Valley and all its tributaries above the lower moraine have wide, flat bottoms (Fig. 4) with a gradual slope upstream. This indi-

cates a large quantity of river gravels as the main aggradational deposit. No lake deposits are anywhere exposed.

Since a moraine of the later glaciation rests on the alluvial fill just above Sunlight basin, it is concluded that the Sunlight lake basin was filled up before the second glaciation occurred. Stream aggradation has continued to the present time, especially above the later moraine.

POSTGLACIAL EROSION

Sunlight Creek has entrenched itself in the earlier moraine. The new valley, lying close to the north side of the wide preglacial valley, has a double profile (Fig. 3), a sharp V-shaped valley about 75 feet deep with 30° slopes cut in the bottom of a broad valley with gentle 10° slopes. On a part of the moraine only the narrow, steep-sided valley is present. Here the valley has entrenched meanders. At the downstream end of this moraine the creek has been superimposed on a spur of the preglacial valley wall and has cut a canyon about 1,000 feet long and 70–200 feet deep in Cambrian limestone.

As the earlier ice melted and retreated, a stream flowed along the north side of Sunlight Valley and quickly cut through the loose material in the higher parts of the moraine. This stream was superimposed on limestone at the downstream end of the moraine. The resistant limestone acted as a local level of baselevel and prevented the stream from cutting very deep in the morainal gravels. However, the stream cut laterally, forming a wide valley in the high part of the moraine, and meandered on the lower parts of the moraine. More recently the limestone spur has been partly cut through, probably by retreat of a waterfall, and Sunlight Creek has again downcut, forming the steep-sided 75-foot valley.

ROCK GLACIERS

Some very remarkable rock glaciers exist down-valley from most of the present glaciers and in some valleys and cirques that have no remaining visible ice. The largest and most important of these is at the end of Sunlight Glacier. This is over ½ mile long and nearly as wide, but the smaller ones are less than 1,000 feet long and a few hundred feet wide. These rock glaciers extend from the cirque headwall or the end of ice down the valley with increasing thickness, to terminate in talus slopes. Their surfaces are very irregular and

usually have narrow concentric ridges and troughs parallel to their sides and lobe-shaped ends (Fig. 5). They are composed of coarse angular fragments averaging ½ foot in diameter and are often underlain by glacier ice.

Identical features occur in the San Juan Mountains of Colorado, where they are called "rock streams." They were first described by Cross and Howe,[8] who considered them to be "rather unusual types of moraines." They were further studied by Howe[9] and by Atwood

FIG. 5.—Rock glacier at the end of Sunlight Glacier

and Mather,[10] who believe them to be a special kind of landslide or rock fall in which the mass is completely broken into debris and transformed into a rapidly flowing body descending streamlike in one sudden rush.

Chamberlin and Salisbury[11] believed in solifluction for the formation of these deposits. This process probably includes (1) the passage of talus over snowbanks; (2) sliding, creeping, and slumping of

[8] Whitman Cross and E. Howe, Silverton Folio, No. 120, *U.S. Geol. Surv. Atlas* (1905).

[9] Ernest Howe, "Landslides in the San Juan Mountains, Colorado," *U.S. Geol. Surv. Prof. Paper 67* (1909).

[10] W. W. Atwood and K. F. Mather, "Physiography and Quaternary Geology of the San Juan Mountains, Colorado," *U.S. Geol. Surv. Prof. Paper 166* (1932), pp. 162–63.

[11] T. C. Chamberlin and R. D. Salisbury, *Geology* (New York: Holt & Co., 1906), Vol. III, p. 474.

bodies of talus perhaps when bound together by ice; and (3) incipient glacial motion.

Capps[12] describes large rock glaciers in Alaska similar to those of the Sunlight area, which he believes form as true glaciers are retreating. Much fractured cirque walls are exposed and undercut as glaciers shrink; this results in a great amount of rock waste being showered down onto the ice, which is carried to the lower edge of the ice and deposited in successive ridges.

The present writer believes that the Sunlight rock glaciers originated as outlined by Chamberlin and Salisbury or by Capps, or by both, because of the following observations: (1) the rock glaciers are always located in cirque basins, (2) several of the rock glaciers are extensions of true glaciers, (3) parallel concentric ridges suggesting successive deposition are always present, (4) talus slopes occur at the ends of the rock glaciers, suggesting slow forward motion, (5) the rock glaciers are concentrated on the north and east sides of the mountains, and (6) interstitial ice is present near the surface of most rock glaciers.

SUMMARY

We may list the following episodes in the Glacial and Recent history of the Sunlight area: (1) early glaciation—moving of ice from the Clarks Fork Valley into lower ends of Absaroka valleys, deposition of older moraine, and formation of lake in Sunlight basin; (2) sedimentation in lake and retreat of ice; (3) cutting of wide valley in moraine and river aggradation of Sunlight Valley and its tributaries; (4) later glaciation—movement of ice down Sunlight Valley and deposition of younger moraine; (5) cutting of narrow valley in older moraine and continued river aggradation in upper Sunlight Valley; and (6) retreat of glaciers and formation of rock glaciers.

ACKNOWLEDGMENTS.—It is a pleasure to acknowledge the financial assistance received from Princeton University for the field work. Thanks are due to Dr. Paul MacClintock for helpful suggestions. M. E. Porter, Raymond B. Koehler, and Donald G. Herring, Jr., served as able field assistants. The writer is under special obligations to Mr. Simon Snyder and others at the Sunlight Ranch for favors and hospitality extended to him during two summers in the field.

[12] S. R. Capps, Jr., "Rock Glaciers in Alaska," *Jour. Geol.*, Vol. XVIII (1910), pp. 359–75.

THE ORIGIN OF THE VALLEY OF JUNE, GULL, AND SILVER LAKES (HORSESHOE VALLEY) MONO COUNTY, CALIFORNIA[1]

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ABSTRACT

This valley, noteworthy for its horseshoe shape in plan, has been variously interpreted as the result of glacial erosion or glacial modification of drainage. It can be shown, however, that neither process was competent. Nor was faulting involved in its formation. The valley is the product of stream erosion and is here explained as the modified meander of a Tertiary river. The stages in its evolution consisted in (*a*) the incision of the meander by lateral and downward erosion, (*b*) further incision by downward erosion alone, (*c*) the isolation of the feature, and (*d*) Pleistocene fluvial and minor glacial modifications. The first three events indicate three relative uplifts of the Sierra Nevada block, which took place during the Pliocene epoch.

INTRODUCTION

The valley of June, Gull, and Silver lakes is located at the eastern foot of the Sierra Nevada, about 10 miles south of Mono Lake (Fig. 1). Its peculiar shape in plan, which Russell[2] aptly compared to a horseshoe, is in striking disharmony with the straightness of the neighboring valleys of this mountain range.[3] No satisfactory explanation accounting in full for this discordant feature can be found in the literature. Another attempt is therefore made to interpret the origin of this valley.

EXPLANATIONS PROPOSED TO DATE

Russell, who first described the valley, devoted no particular section to its origin. His remarks bearing on the question are scattered through an account of Pleistocene Rush Creek glacier. They strongly suggest an origin through glacial erosion, although Russell carefully avoided a specific statement to this effect, as he did not believe in great competence of glacial erosion. His most definite suggestion

[1] Read before the Association of Pacific Coast Geographers, June 29, 1939.

[2] Israel C. Russell, "Quaternary History of Mono Valley, California," *U.S. Geol. Surv. 8th Ann. Rept., 1886–87*, Part I (1889), p. 342.

[3] This morphologically unique valley has not yet been named. Making use of Russell's fitting description, the name of Horseshoe Valley is herein applied to it.

invokes a glacial modification of drainage consisting in a reversal of stream flow in the eastern arm of the present-day horseshoe. This

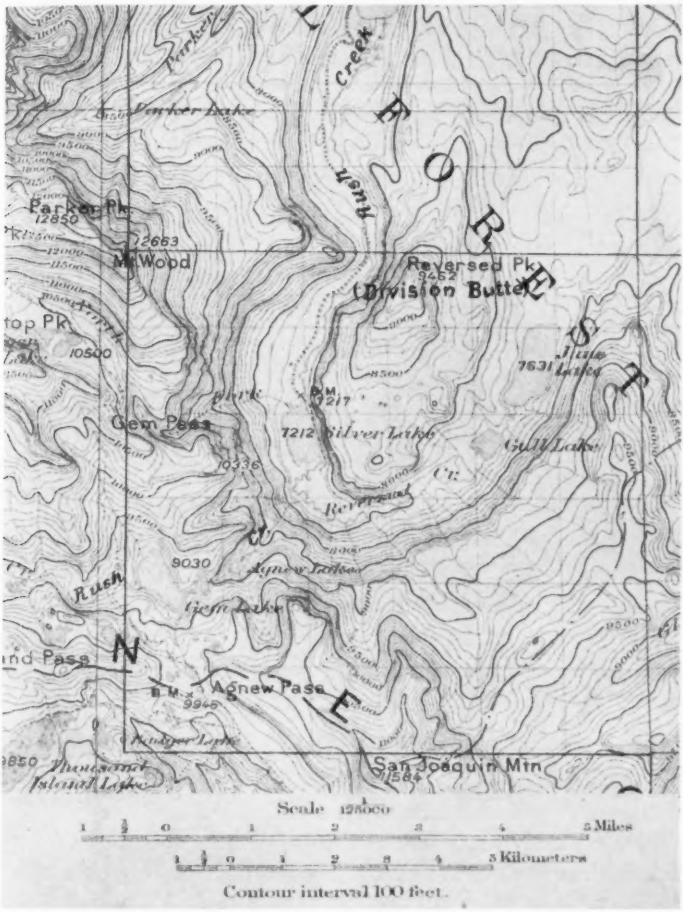

FIG. 1.—The horseshoe-shaped valley of June, Gull, and Silver lakes. (From U.S. Geological Survey, Mount Lyell quadrangle, California.)

interpretation implies that the feature was started by two creeks which, from sources near the toe of the horseshoe, once drained

northward and northeastward, respectively. A series of low terminal moraines, at most 200 feet high, located at the north end of June Lake (Fig. 2), is credited with having caused this reversal of drainage in the eastern arm.[4] In order to facilitate the obviously tremendous erosive task involved in the postulated reversal of drainage, a de-

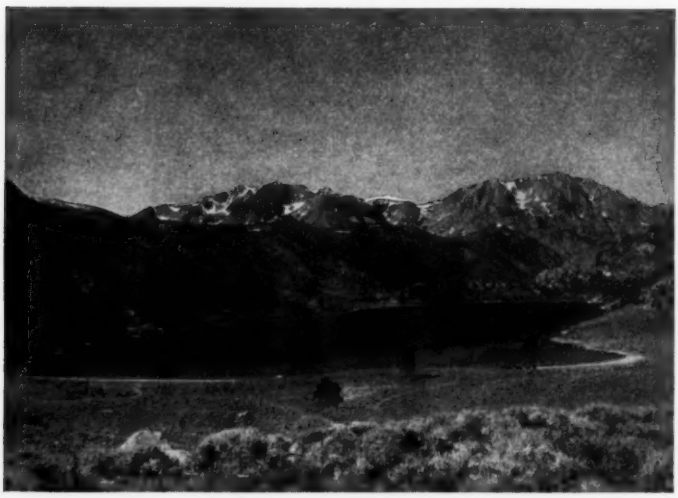

FIG. 2.—June Lake seen from the north. At its proximal end, moraines of Rush Creek glacier of the Tioga (late Wisconsin) stage, which are responsible for its present northern limit but not for the damming of the lake. At the distant end of the lake, the low granitic divide behind which Gull Lake is situated. The lake has no outflow channel.

tached body of ice, lingering in the basin of present-day June Lake during the early stages of deglaciation, is conveniently called upon by Russell to provide an abundant source of water.[5]

Salisbury and Atwood followed Russell in his assumption that the drainage was reversed and that a detached body of ice once occupied the site of June Lake, but credited the reversal of drainage to the damming of runoff by this detached body of ice.[6]

[4] Russell, *op. cit.*, p. 343. [5] *Ibid.*, p. 368.

[6] Rollin D. Salisbury and Wallace W. Atwood, "The Interpretation of Topographic Maps," *U.S. Geol. Surv. Prof. Paper 60* (1908), p. 76.

CRITICAL VALUATION OF THESE EXPLANATIONS

These various ideas call for scrutiny. The theory of glacial origin of Horseshoe Valley appears substantiated by its good U-shaped cross section (Fig. 3). But no glacier ever passed through the valley in the manner suggested by its trough form, i.e., entering at one end and issuing from the other. Pleistocene Rush Creek glacier, which

FIG. 3.—View of the glaciated trough at the toe of Horseshoe Valley. Silver Lake in the middle distance. To the right of center, the hanging valley of upper Rush Creek, which provided nearly all the ice of Pleistocene Rush Creek glacier.

repeatedly filled the valley with ice, entered it from the southwest at the toe. Forced to separate by Division Butte,[7] it sent a branch into each arm of the horseshoe. If effective glacial erosion be admitted, the resulting form would consist of two glacial troughs, corresponding to the two arms of the Pleistocene glacier, separated by a ridge running southwestward from Division Butte and of a height commensurate with the depth to which the flanking troughs had been eroded. Instead we find a continuous circular trough. The toe of

[7] Reversed Peak of the U.S. Geological Survey Mount Lyell quadrangle. The older name, given by Russell, appears to the writer to be more appropriate.

the horseshoe, trending at right angles to the direction of flow of the ice, obviously cannot be explained by processes of glacial erosion. The catchment basin of Rush Creek is a hanging valley with respect to Horseshoe Valley (Fig. 3). Usually a potent argument in favor of glacial erosion, this relationship is here, on the contrary, an argument favoring inefficiency of glacial erosive processes, for the ice which issued from this hanging valley was not tributary to a larger glacier but was the main glacier of the area. The step below Gem Lake, nearly 2,000 feet high, over which Rush Creek glacier cascaded into Horseshoe Valley must thus be of nonglacial origin. As Pleistocene Rush Creek glacier ended in two separate tongues at a very early stage of glaciation, Division Butte must date back to early glacial or even preglacial time, necessitating a similar age for Horseshoe Valley.

Russell's assumption that the valley resulted from a reversal of drainage affecting its entire eastern arm lacks any foundation. A reversal of drainage really occurred here during the Wisconsin stage of glaciation, but it affected only the northernmost part of the eastern arm, the area now occupied by June Lake. A bathymetrical map of the lake by the Department of Water and Power of the City of Los Angeles shows that it resulted from the damming of a formerly northward-draining valley. The former divide is still recognizable in the low granitic spur between June and Gull lakes (Fig. 2). As the overflow of this body of water took a southwesterly route, it must have found a gradient in this direction, since a relatively small further rise of the lake would have resulted in an overflow across the low terminal moraines at its north end. The granitic divide indicates the existence of a southwestward-draining valley in the location of present-day Reverse Creek[8] at a time preceding the formation of June Lake. The insignificant and infrequent overflow of this lake, a consequence of the small size of the area tributary to it, cannot be credited with much erosive power and could not possibly have carved the valley of Reverse Creek to a depth of over 300 feet since the last

[8] The name Reverse Creek, suggesting only an abnormal direction of flow, is held to be preferable to the explanatory but mistaken form of Reversed Creek used by Russell and rendered official by usage on the U.S. Geological Survey Mount Lyell quadrangle.

glaciation, as implied by Russell. In the same time much larger creeks in the neighborhood have accomplished only an insignificant fraction of the work which would have to be credited to Reverse Creek if a reversal of its course had occurred. Russell's other postulate, that the melting of a detached body of ice occupying the site of June Lake provided the necessary runoff for the carving of the valley of Reverse Creek, helps little, as it would require that meltwater remove a volume of bedrock corresponding to about half the volume of the detached body of ice. Furthermore, before this meltwater could have started its work of reversing the drainage, Rush Creek glacier would have had to evacuate Horseshoe Valley by retreating to the hanging valley above Silver Lake. Under such conditions, so obviously unfavorable for the preservation of ice, it is impossible to visualize a relatively thin detached body of ice surviving at the site of June Lake while the main glacier in its retreat lost at least half of its volume. The hypothetical detached body of ice is obviously an impossibility. The same difficulties confront the explanation of reversal advanced by Salisbury and Atwood. Well-preserved, though low, recessional moraines of the last glacial stage are encountered in the upper half of the valley of Reverse Creek. They would have been completely removed if a reversal of drainage had occurred there since the last stage of glaciation. Their preservation proves that Reverse Creek has today essentially the same gradient it had prior to the Wisconsin stage of glaciation.

The distribution of moraines of Pleistocene Rush Creek glacier shows conclusively that throughout the second half of the glacial epoch the western branch of the glacier was considerably larger than the eastern branch. But as the eastern arm of the horseshoe is the wider and, furthermore, lies in the direct extension of the hanging valley from which the ice issued, one would naturally expect its glacier to have been the larger of the two. That such was not the case suggests impeded glacial flow in the eastern arm which indicates the presence of a powerful barrier in this arm, obviously the divide at June Lake. This in turn necessitates a southwestward-draining valley at the location of present-day Reverse Creek throughout the latter part of the Pleistocene epoch—conditions of drainage as they are encountered today. Horseshoe Valley thus becomes definitely a

nonglacial feature essentially completed at an early stage of, or even prior to, the Pleistocene glaciation.

INFLUENCE OF FAULTING

As faulting left marked imprints in the morphology of the region, one is tempted to ascribe the formation of Horseshoe Valley to structural processes. The valley can be visualized as a graben, bounded

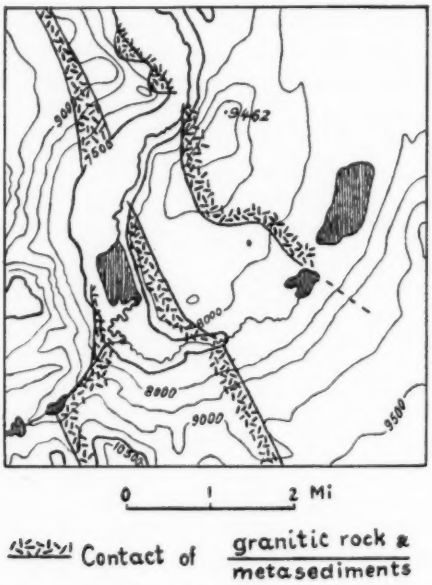

FIG. 4.—Geologic sketch map of Horseshoe Valley

by two nearly circular faults, or it can be assumed to follow a single fault bounding the circular block of Division Butte. A relative rise or drop of the latter, or an overthrusting toward the southwest or a heave toward the northeast, would aid materially in explaining this feature. Any tectonic interpretation read out of the topography of the region, however, fails before other solid evidence. No trace of faulting can be found in either arm of the horseshoe, whether on its flanks or in its bottom. Convenient datum planes which would reveal displacement by faulting are provided by the two contacts of

granites and metamorphic rocks which cross both arms of the valley from the northwest to the southeast (Fig. 4). The northern gives decisive evidence where it crosses the western arm. The contact plane, easily recognizable on either slope, dips northeastward at a high angle but is not vertical. Viewed from Division Butte the two parts of the contact to the west and the east of Rush Creek can be seen to lie still in the same plane. No displacement, therefore, either vertical or horizontal, has occurred here. The southern contact is well exposed in the eastern arm of the horseshoe and similarly crosses it without any offset. Horseshoe Valley, therefore, cannot be explained by tectonic processes.

SCULPTURE BY RUNNING WATER

As neither faulting nor glacial erosion was responsible for the formation of Horseshoe Valley, stream erosion remains the only competent process. The nearly circular shape of the valley suggests an entrenched meander. Division Butte answers well to the characteristics of a meander spur, its gentle southeastern side corresponding to a slip-off slope, the steep northwestern and eastern sides to undercut slopes. From this it is deduced that the direction of flow of the river was from the western arm through the toe into the eastern arm—a drainage opposite that of today. The source of the river must thus have lain somewhere to the north, seemingly at a considerable distance, as the diameter of the meander indicates a river of at least the size of the Cumberland River at Monticello (Kentucky).

The meander-spur shape of Division Butte ends at about 8,300 feet. Below this level lies the trough of Horseshoe Valley proper, 600 feet deep at June Lake, 1,000 feet deep in the entire western arm. It appears logical to credit the same river with its formation and to attribute the efficiency in downcutting to a strong rejuvenation of erosion. The topography indicates, however, that the erosive behavior of the river suddenly changed from a combination of downward and lateral erosion to downward erosion alone. Down to the 8,300-foot level the stream obviously governed the form, increasing the diameter of the meander through lateral erosion to conform to the conditions of volume and gradient. But below this level the me-

ander remained unchanged in size, thereby indicating that lateral erosion was completely eliminated. Rejuvenation of erosion would not stop lateral corrasion but would accelerate it, as well as downcutting. The lateral component of erosion, however, would decrease in importance as the downward component increased in efficiency but would still force an increase in size upon an incised meander. The complete absence of lateral erosion in the formation of Horseshoe Valley therefore demands an explanation other than rejuvenation. As the volume of water of a river determines the size of its meanders and the amount of lateral erosion in an incised meander, it is here concluded that the elimination of lateral erosion was due to a decrease in discharge of the river of Horseshoe Valley. Climatic changes might be called upon to explain the dwindling of the river. But as the reduction in volume of the river was combined with increased erosive power, a relative rejuvenation through orogenic processes which simultaneously caused the loss of part of the catchment area of the river appears more probable, especially as the area has been tectonically unstable from the Tertiary to the present. Since all tributary valleys, even that of upper Rush Creek, were left hanging by the rapid downcutting of the reduced river, it is apparent that its volume still remained large. As a large volume could hardly have been retained in a reversal of drainage, it follows that the direction of flow remained as before, i.e., opposite to the present direction of drainage.

Horseshoe Valley, once part of an extensive stream system, is at present isolated. Tertiary and Quaternary lavas and tuffs, as well as Pleistocene glacial deposits, conceal its northward extension. As the valley ends at the edge of the bedrock complex of the Sierra, a relative uplift along the Sierra fault must be responsible for the separation. After uplift the valley was taken over as main drainage channel by the runoff of the surrounding area, which in time established the present stream system. Since the drainage now goes from east to west, the same direction of drainage must have existed in the raised and deserted meander. The uplift was thus accompanied by a tilting toward the northwest, which originated a reversal of drainage. The northern end of the eastern arm, now occupied by June Lake, either retained the original direction of drainage or was captured through

retrograde erosion of a stream to the north. It drained northward until Pleistocene lavas and moraines interfered and caused the formation of June Lake. At present the remaining part of Horseshoe Valley varies in gradient, which is steep, over 100 feet to the mile, in the sector drained by Reverse Creek, but gentle, 5 feet to the mile, in the sector traversed by the master stream, Rush Creek. This difference in gradient cannot have existed immediately after uplift and tilting of the meander, which had an even gradient. It has resulted from fluvial and glacial erosion since the uplift. As the gradients are in marked disharmony with the direction of glacial flow, in the toe as well as in the eastern arm, but in good harmony with the volume of discharge of the creeks in the valley, they demonstrate that the last modifications are mainly the result of stream erosion. The trough shape, often thought to be exclusively of glacial origin, must therefore be admitted to have been created here by stream erosion, slope-weathering, and alluviation. The elevation of the spur between June and Gull lakes, the former divide, indicates that 600 feet of the depth of Horseshoe Valley have to be credited to the erosive action of the reduced river passing through the meander. The remaining deepening, negligible at this divide but amounting to 400 feet in the western arm, therefore represents the work of Rush and Reverse creeks, accomplished since the isolation of the meander.

APPROXIMATION OF AGE

It remains to establish the comparative age of the different events involved in the formation of Horseshoe Valley. Since fossiliferous Tertiary sediments are lacking, the determination must depend upon morphologic criteria. In the Yosemite Valley, 25 miles to the west, Matthes[9] recognized three stages in its evolution: the broad-valley, the mountain-valley, and the canyon stages. Based upon physiographic similarity, the valley of the incised meander must be correlated with the mountain-valley stage, while Horseshoe Valley proper appears to correspond to the canyon stage of the Yosemite. According to the age determination established by Matthes for the Yosemite, the formation of the incised meander was "initiated . . . . at

[9] François E. Matthes, "Geologic History of the Yosemite Valley," *U.S. Geol. Surv. Prof. Paper 160* (1930).

the end of the Miocene epoch . . . . and lasted presumably through most if not all of the Pliocene epoch," while Horseshoe Valley proper would have evolved "in all probability wholly during the Quaternary period."[10] Since it has been shown that Horseshoe Valley had nearly reached its present configuration at least at an early stage of glaciation, if not actually prior to this event, its formation can hardly be assigned wholly to the Quaternary. In my opinion all stages in the evolution of this valley, from the incision of the meander to the reversal of drainage in the abandoned meander, occurred during the Pliocene epoch.

Instead of the two uplifts indicated in the Yosemite by its mountain-valley and canyon stages, three relative uplifts are evident in Horseshoe Valley, viz., (*a*) an uplift leading to the evolution of the incised meander, (*b*) an uplift initiating Horseshoe Valley proper through the very active downcutting of the river of reduced volume, and (*c*) an uplift which isolated the feature. The last two correspond to the one uplift manifest in the canyon of the Yosemite. Were it not for the isolation of the meander, the two could not be differentiated in Horseshoe Valley. Matthes' postulated single uplift initiating Yosemite Canyon may thus well consist of two successive uplifts.

That the region surrounding Horseshoe Valley has not remained quiescent since the last major uplift of the Pliocene epoch is indicated by small fault escarpments traversing moraines of Pleistocene Rush Creek glacier. They indicate small-scale local faulting in the vicinity of the Sierra Nevada fault which occurred at four different periods of the Pleistocene.

[10] *Ibid.*, p. 1.

A PHYTOSAUR IN UNION COUNTY, NEW MEXICO WITH NOTES ON THE STRATIGRAPHY

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ABSTRACT

A skull of the phytosaur *Machaeroprosopus*, discovered in the Sloan Canyon formation, confirms the Triassic age of the rocks lying beneath the Exeter (Jurassic) in the valley of the dry Cimarron River of Union County, New Mexico, and Cimarron County, Oklahoma. The underlying red beds, the Sloan Canyon formation, and the Sheep Pen Canyon formation form a continuous and conformable succession of strata in this region, and the conclusion is drawn that these three units should be included in the Dockum group.

MATERIAL AND LOCALITY

A skull of *Machaeroprosopus* sp. indet. with partial lower jaws and associated postcranial elements (University of Oklahoma Museum of Geology and Paleontology, No. 1250) was found by the junior writer on the north side of Sloan Canyon Creek 200 yards downstream from the crossing of U.S. Highway 64, in Union County, New Mexico, 11 miles west of Kenton, Oklahoma. The fossil horizon is in the Sloan Canyon formation of Triassic age, about 100 feet below the top of the formation.

DESCRIPTION

The specimens obtained were not more than 1 foot below the surface of the ground and were badly cracked. Compaction within the containing deposit had flattened the skull, and its posterior part was rolled under toward the left side. Although there is considerable fracture and distortion in this skull, it is the conviction of the writers that the specimen shows enough distinctive characters for generic determination.

The skull (Figs. 1 and 2).—The maximum length of the skull is 1,170 mm. and the postnarial length, measured from the anterior border of the nares, is 38 per cent of its total length. The occiput is

elevated and bears a surface sculpture of many pits and anastomosing grooves. This rugosity is heavy along the posterodorsal portion, is faint on the premaxillaries, but becomes distinct again on the downturned tip of the rostrum.

The rostrum is evenly tapered and bears a slight mid-dorsal premaxillary swelling, which is most acute at a point above the lateral

FIG. 1.—*Machaeroprosopus* sp. skull, dorsal view. Total maximum length, 1,170 mm.

FIG. 2.—*Machaeroprosopus* sp. skull, lateral view. Total maximum length, 1,170 mm.

union of the premaxilla and maxilla and forms a prenarial crest or hump. Anterior to the hump, the rostrum is slender. Posterior to the premaxillary hump, the nasals form a prominent cratered dome. Camp[1] regards the occurrence of this feature in the genus *Machaeroprosopus* as a female character. The narial crater is located in an extremely posterior position, its posterior border being only 80 mm. in front of the anterior border of the orbits. The nares are large and compressed, their outer walls being convex toward the mesial line. The portion of the narial septum preserved in this skull is relatively thick.

[1] C. L. Camp, "A Study of the Phytosaurs with Description of New Material from Western North America," *Mem. Univ. Calif.*, Vol. X (1930), p. 28.

The orbit is somewhat D-shaped, with the flattened side at the inner margin. The greatest longitudinal diameter of the left orbit is 89 mm., while the greatest transverse diameter is 78 mm. A prominent ridge bounds the orbits in the prefrontal frontal region.

The dentition is distinctly heterodont. Two longitudinal ridges on the ventral side of the rostrum lie mesiad of the tooth rows, features that have been previously defined as buffers against the lower jaw to prevent the teeth from being broken when the jaws were forcibly closed.

Of the eighteen primitive phytosaurian characters listed by Camp,[2] this skull shows but two: rostrum not exceptionally long; auditory meatus a notch. Three other characters are indeterminable on this specimen: the length of the interpterygoid vacuity; the size of the postpalatine fenestrae; the relationship of the pterygoid to the ectopterygoid.

In Table 1 the various skull characters of the phytosaur from Union County are compared with three species of *Machaeroprosopus* collected by the University of California in Arizona. It may be noted from this comparative chart that the University of Oklahoma animal is quite similar to the young *Machaeroprosopus tenuis* Camp, the former being larger in all characters at a constant ratio. However, the writers think it unwise to refer the University of Oklahoma specimen to *M. tenuis* until additional material has been studied.

The lower jaw.—Only the anterior half of the lower jaw was recovered. It contains one anterior dentary tooth which is very similar to the enlarged teeth in the anterior tip of the premaxillaries. Like the skull, the lower jaw shows the two longitudinal "buffer" ridges and also the alveoli separated by bony septa. Sutures are indistinguishable, and all other characters are obscured or missing.

The dermal scutes.—Numerous dermal scutes were found in close association with the skull and lower jaws. These scutes are typically rugose but are not pitted so deeply as the scutes of *Angistorhinus alticephalus* Stovall and Wharton in the University of Oklahoma collection. The scutes differ in shape and size. The smallest one obtained is elliptical in outline and measures 10 × 16 mm.; the largest is

[2] *Ibid.*, p. 148.

TABLE 1

SKULL CHARACTERS IN THE SPECIES OF *Machaeroprosopus*

	M. tenuis Univ. Calif. Mus. Pal. No. 27018	M. lithodendrorum Univ. Calif. Mus. Pal. No. 26688	M. adamanensis Univ. Calif. Mus. Pal. No. 26699	Univ. Okla. No. 1250
Total length of skull	819 mm.	1,215 mm.	1,095 mm.	1,170 mm.
Ratio of postnarial length to total length of skull	37 per cent	46 per cent	42 per cent	38 per cent
Mid-dorsal premaxillary swelling	Placed far forward in adult	Obliterated by high, elongate rostral crest	Slight	Slight
Relative width of postorbital squamosal bar	Broad	Narrow	Narrow	Broad
Rostrum	Long, massive, no crest in adult female	Short, massive, elongated crest	Long and slender, short crest	Long and slender, no crest in adult female (?)
Two enlarged teeth on tip of premaxillary	*	*	*	*
Teeth on one side of upper jaw	42–45	44–45	40–43	43–45
Posterior maxillary teeth enlarged, short, compressed, serrate	*	*	*	*
Diameters of antorbital fenestra	Long, 85 mm.	Long, 204 mm.	Long, 161 mm.; short, 57 mm.	Long, 130 mm.; short, 47 mm.
Anteroposterior diameter of orbit	62 mm.	100 mm.	87 mm.	89 mm.
Ratio of length of antorbital fenestra to diameter of orbit	1.37	2.04	1.85	1.46
Excavation of supratemporal fenestra	Narrow and to depth of only fourth of parietal	Broad and to depth of third of parietal or less	Broad and to depth of nearly half the length of parietal	Like *tenuis*

* This symbol indicates full possession of the character stated.

subtriangular in outline and measures 104 × 106 mm. All the scutes associated with this animal possess a central comb or ridge, more prominent on the smaller scutes than on the larger ones.

STRATIGRAPHY OF THE BEDS

Fossil fragments were found at two main levels in a series of shales, calcareous sandstones, and marls. This series of beds was designated the "Sloan Canyon formation" by Parker.[3] The entire outcrop of Triassic rocks at this place in Sloan Canyon totals more than 100 feet in thickness, but only the lower 50 feet were measurable at the fossil location. The beds dip generally to the southeast, but minor folds lend complexity to the structure. The locality from which the fossils considered in this paper were taken is 100 yards downstream from a clastic dike system shown on a map by Parker[4] about 2 miles south of the northwest corner, T. 31 N., R. 36 E. Table 2 shows the stratigraphic section at the collection site.

Stovall gives a 67-foot section of the Triassic exposed 11 miles east of the foregoing location at Tate Butte, 4 miles north of Kenton, Oklahoma. In this section are thin, alternating beds of purple to gray limestones; buff, bluish-gray, and gray sandstones; and brick-red to gray shales. He says: "The Triassic exposures in the western end of the region are typical. The color is purplish red mottled with gray. In some places, however, gray becomes the dominant color and the purple mottling is secondary."[5] Correlation may be made between the Sloan Canyon formation and the section shown by Stovall, for the Triassic rocks may be traced easily along the floors of the Sloan and Rio Cimarron canyons from the phytosaur locality into Oklahoma.

Since the Dockum group is a series of continental deposits and is irregular and lenticular, determination of time equivalents within this group over a wide area is difficult. Sellards, Adkins, and Plum-

[3] Ben H. Parker, "Notes on the Occurrence of Clastic Plugs and Dikes in the Cimarron Valley Area of Union County, New Mexico," *Kan. Geol. Soc. Fourth Ann. Field Conf.* (1930), pp. 131–37.

[4] "Clastic Plugs and Dikes of the Cimarron Valley Area of Union County, New Mexico," *Jour. Geol.*, Vol. XLI (1933), p. 39.

[5] J. Willis Stovall, "The Geology of Northwestern Cimarron County, Oklahoma" (Doctor's thesis, University of Chicago, unpublished [1938]), p. 7.

mer[6] reviewed the literature on the Triassic of Texas and proposed a correlation of the various units of the Dockum previously set forth. Table 3 indicates their correlation.

Parker[7] described two new Triassic formations exposed in the valley of the Rio Cimarron lying unconformably beneath the Exeter sandstone (Jurassic). One of these, the Sloan Canyon formation, lies above the typical red sandstones and shales of the Dockum group and is a series of varicolored shales interbedded with hard gray marl

TABLE 2

	Feet
Talus-covered slope	49
Purple-red shale, interbedded with thin, gray to green calcareous sandstones	15
Soft, purple-maroon shale, splotched gray or green, containing *Machaeroprosopus*	4
Hard, gray to maroon marl	½
Purple-red shale splotched with gray or green	10
Gray to purple marl, portions of which are composed largely of *Unio* and unidentified gastropods	6
Red shale spotted with gray and green	10
Hard gray calcareous sandstone	2
Soft red shale	3
Friable conglomerate (pebbles of marl or limestone) red in color	1
	100½

and a bed of red sandstone near the base. Conformably overlying the Sloan Canyon formation is the Sheep Pen Canyon formation, a tan to buff-colored sandstone, thin-bedded to massive. The writers visited the type locality of the Sheep Pen, as well as the Sloan Canyon formation, and are convinced that Parker was justified in distinguishing these two formations.

The contact of the predominantly red-colored strata of this area with the overlying variegated shales of the Sloan Canyon formation is generally concealed by debris moved down the hillsides from the higher formations. However, a 2–3-foot layer of conglomerate immediately overlies the red-colored beds on the canyon wall ¼ mile

[6] E. H. Sellards, W. S. Adkins, and F. B. Plummer, "Geology of Texas," *Univ. Texas Bull. 3232* (1932), p. 243.

[7] "Clastic Plugs and Dikes . . . . ," *op. cit.*, pp. 38–52.

southeast of Battleship Mountain, a small butte in Sec. 28, T. 32 N., R. 35 E. This conglomerate is believed to be a basal bed in the Sloan Canyon formation. The parallel relationship of this conglomerate with beds above and below it and the limited extent of the bed indicate that it represents, at most, only a very slight and local change in depositional sequence over the area. Figure 3 is a generalized cross section from Battleship Mountain to the small butte at the type locality of the Sheep Pen sandstone.

Parker[8] considered the Sloan Canyon and Sheep Pen Canyon formations to be Triassic rocks definitely above the Dockum group.

TABLE 3

Eastern New Mexico (Darton)	Southern Panhandle of Texas		Central Panhandle of Texas (Drake)	Northern Panhandle of Texas (Gould)
	(Adams)	(Hoots)		
Chinle shales	Chinle shales	Upper red clay	Sandy clay, some sandstone	(Thin or absent)
"Santa Rosa" sandstone	"Santa Rosa" sandstone	Basal red clay and sandstone	Sandstone and conglomerate, some clay	Trujillo sandstone and shale
(Generally absent)	Basal shales	(Generally absent)	Sandy clay	Tecovas basal shale

This correlation was made on the basis of lithologic characters and relative stratigraphic position without aid of fossil evidence. The Sloan Canyon and Sheep Pen differ greatly in color and lithology from any other exposed Triassic rocks throughout New Mexico and Texas. In these two states the Triassic is composed chiefly of red shales alternating with red to gray sandstones. Notwithstanding the fact that the Sloan Canyon formation is a distinct lithologic unit, there are no facts at hand to prove that it might not be a time equivalent of upper members of the Dockum elsewhere. The presence of *Unio* and *Machaeroprosopus* in the Sloan Canyon strongly suggests Dockum age, because both of these genera have been found extensively in the Dockum. The Triassic of Union County is isolated from other beds of the same age to the south and southeast

[8] "Clastic Plugs and Dikes . . . . ," *op. cit.*, p. 40.

in New Mexico and Texas by overlapping sedimentary and igneous rocks of Tertiary and Quaternary age. This overlapping belt of Tertiary and Quaternary is not less than 50 miles wide, a sufficient distance for marked gradations in lithology of the underlying Triassic rocks. There is a possibility that the Sloan Canyon–Sheep Pen beds in the region under consideration might be, in part, equivalent to the Trujillo–Santa Rosa members of the Dockum or, in part, to the Chinle shales member. Since *Machaeroprosopus* is also found in the Chinle formation of Arizona, it is also likely that the Sloan Canyon–

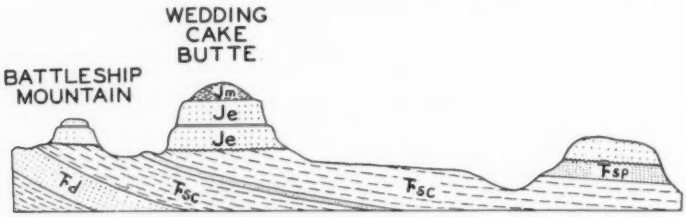

FIG. 3.—Diagrammatic cross section in Union County, New Mexico, showing the angular unconformity between the Jurassic and the Triassic formations: typical Dockum "red beds," *Trd;* Sloan Canyon formation, *Trsc;* Sheep Pen Canyon formation, *Trsp;* Exeter sandstone, *Je;* and Morrison formation, *Jm.*

Sheep Pen are correlatives of that formation. The writers are fully aware, however, that vertebrate and invertebrate genera are not reliable standards for a detailed correlation of horizons over an area as extensive as that covered by the Triassic of New Mexico, Texas, and Arizona.

CONCLUSION

The writers think it necessary to designate the phytosaur discovered in Union County, New Mexico, as *Machaeroprosopus* sp. indet. The discovery of this genus substantiates previously stated opinions that the Sloan Canyon formation is Triassic in age. Sloan Canyon and Sheep Pen Canyon are formations distinctly different from other Triassic rocks of New Mexico and Texas. Notwithstanding difference in lithologic characters, the entire section of Triassic in Union County should be included in the Dockum group, a conclusion supported by fossil evidence.

AN OCCURRENCE OF TINSTONE IN THE PRE-CAMBRIAN OF WESTERN ONTARIO

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ABSTRACT

At Eagle Lake in western Ontario are early pre-Cambrian metamorphic rocks intruded by two different granites. The earlier, a gray granite, has dikes of light-colored pegmatite, which in one locality take the form of a tinstone-bearing pegmatitic greisen. A tinstone-bearing greisen has also been reported from eastern Manitoba, and a reddish pegmatite at Shatford Lake in eastern Manitoba also carries tinstone. The several occurrences indicate a tin-bearing area of about 150 miles in length.

During the summer of 1936 the writer was employed in a geological examination of a map-sheet area which included the central part of Eagle Lake—a body of water situated in western Ontario just south of the Canadian Pacific Railway and about 100 miles from the western boundary of the province. The rocks are entirely pre-Cambrian except for glacial and recent deposits. Those with which this paper deals comprise early Archean metasedimentary gneisses and metamorphosed volcanics, intruded first by a gray and later by a red granite. Each of the granites has a suite of pegmatite dikes, which are white in the case of the gray and red if they accompanied the younger red granite.

On one of the many arms of Eagle Lake known as Partridge Bay, which lies a few miles from the Vermilion Bay station on the railway, our party discovered very narrow dikes of white pegmatite, corresponding in appearance to the pegmatite associated with the older gray granite, but differing in general composition. It agrees perfectly with a pegmatitic type known as "greisen" in the tin fields of Cornwall. It consists mainly of granular quartz with lesser amounts of feldspar in perfectly bounded crystals and of biotite, muscovite, and cassiterite. A small amount of tourmaline is also present. The cassiterite (tinstone) is distinguished easily from the other minerals except tourmaline. From tourmaline it is distinguished macroscopically by the adamantine luster of its crystals as compared with the glassy appearance of tourmaline, and by square prisms striated transversely, as contrasted with the triangular prisms of tourmaline

which are longitudinally striated. The color of the powder, brown for tinstone and white for tourmaline, is perhaps the most valuable property for macroscopic discrimination in the field. In thin sections under the microscope tinstone shows light- to dark-brown pleochroism often with bands somewhat resembling the twinning of plagioclase under crossed nicols, but less sharply defined and of a different orientation; it is also well distinguished by the metallic appearance and high relief of the transparent section by reflected light when the light from below the stage is shut off, and by its optically positive character as determined by the interference figure. Tourmaline has a different pleochroism and is optically negative.

This occurrence of tinstone is probably the first found in the province of Ontario. Specimens of greisen, very similar in appearance, which were said to have been found some years ago in Manitoba, near the eastern boundary, have been shown to the writer by reliable parties. Another occurrence of tinstone at Shatford Lake, Manitoba, is of quite a different character. A dike of reddish pegmatite has large crystals of tourmaline arranged in parallel at right angles to the strike of the dike. Imbedded in these tourmaline crystals are masses of cassiterite almost indistinguishable to the eye but readily differentiated by their brown powder when scratched. The color and type of the feldspar would suggest that this pegmatite is of a different age from that which produced the Eagle Lake greisen. Tin has also been found by De Lury on the west side of the Lake of the Woods in a mineral which he thought to be smaltite.

During exploratory work from Lac Seul northeastward to Birch Lake in the district of Patricia, Ontario, in 1919, the writer encountered tourmaline-bearing binary granite areas like those which are typical of tinstone-producing countries in other parts of the world. He found tourmaline in the associated quartz veins but no tinstone. This area, like the others mentioned in this paper, has been productive of gold in more or less abundance in several of its parts.

Taken altogether, these mineral occurrences indicate a tin-bearing area about 150 miles long on a southeast to northwest axis in the provinces of Ontario and Manitoba, while indications of possible tinstone occurrences are found for about the same distance to the northeast at a right angle to this axis.

ON THE CRYSTALLIZATION OF THE KANGASNIEMI ORBICULAR GRANITE

D. S. BELYANKIN AND V. P. PETROV
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Pentti Eskola in his recent paper "On the Esboitic Crystallization of Orbicular Rocks"[1] objects to the common conception of the crystallization from the fluid magma of the orbicular elements of these rocks. He speaks in this connection as follows:

> For the "true esboites" application of this idea meets with difficulties: (1) the amount of albite does not increase from orbicule to matrix or does so less than should be expected; (2) the sequence of the zones of orbicules and the occurrence of pegmatitic nuclei are often contradictory to the laws of magmatic crystallization; and (3) many examples of this group are isothrausmatic, having portions of the matrix as nuclei. The power of evidence, though perhaps not quite decisive, is decidedly in favor of the idea that the matrix was solid, or at least contained a solid sponge of early-formed crystals, at the time of the crystallization of orbicule.[2]

Having read Eskola's paper, we were glad to note that this author's general conception concerning the relatively late crystallization of the orbicules of esboite agrees with our personal observations of the most important occurrence of esboites, viz., the orbicular rock from Kangasniemi. However, it is a great pity that our corresponding article published in 1933[3] remained apparently unknown both to the writer and to the editor of the *Journal of Geology*. Therefore, we think it timely to beg the editor to reproduce on the pages of his journal the following short extract from our paper.

Owing to our investigation, supplementary to Frosterus' data, we are able to state the following facts:

1. The nucleus of the sphere may contain not only the fragments of gneiss but also granite identical with that in the interstices between the spheres.

2. The plagioclase zones in the orbicule shells by the relative

[1] *Jour. Geol.*, Vol. XLVI (1938), pp. 448–85. [2] *Ibid.*, p. 483.

[3] *Travaux de l'Inst. Petrographique Acad. Sci. U.S.S.R.*, Book III (1933), pp. 21–37.

content of colored components and by the character of plagioclase are intermediate between the granite and the microcline zone.

3. The microcline zone predominates in the concentric shells by its volume. It has the following characteristic features: (*a*) it is relatively leucocratic, (*b*) its plagioclase is more acid in comparison with that in the interstitial granite, and (*c*) it has an aplite-micropegmatitic structure consisting of the radii of microperthite pierced by the plagioclase.

TABLE 1*

QUANTITATIVE MINERALOGICAL COMPOSITION OF SEPARATE STRUCTURAL ELEMENTS OF THE KANGASNIEMI ORBICULAR ROCK

	1	2	3
Quartz	32.9	32.1	39.3
Microcline-perthite	34.3	56.7	17.7
Plagioclase	20.7	9.9	35.7
Per cent of An in plagioclase	30.0	20.0	28.0
Biotite	12.1	1.3	7.3

* Values in weight per cent.

1. The granitic intermediate rock of the orbicules.
2. The microcline zone.
3. The plagioclase zone.

The quantitative mineralogical composition calculated microscopically (and recalculated in weight per cent) is represented in Table 1.

Considering these facts, we assume, in contrast with Frosterus and Sederholm, the normal sequence of crystallization of both chief components of the Kangasniemi rock: first the granite of the nucleus and that of the interstitial mass and then the aplite-pegmatite of the microcline zone.

This normal succession of crystallization was, however, accompanied by its anomalous structure development. Instead of filling up the fissures and other hollows in the granite body, the aplite-pegmatite magma forms here orbicules, a concretionary form in the very body of granite. In order to explain this peculiarity we must look for some special cause, which is usually absent in the crystallization of the granitic magma. Considering some structural

analogues, we are inclined to see this cause in the increased contact phenomena, in abundant contact gaseous emanations, which partly dissolved granite and regenerated it in the form of aplites and pegmatites. Water above the critical temperature must have taken the greatest part in these emanations, but probably there were dissolved also some other substances, e.g., potassium compounds in which the orbicular rock is rich compared with the normal granite. It is possible that, owing to the gas pressure from the reaction of gaseous emanations with granite, mechanical fracturing also occurred. The fragments thus produced served afterward as the centers of crystallization of orbicules.

The zonal arrangement of shells, their differentiation into the plagioclase and microcline zones, might also arise from the unequal stability in contact with the gaseous emanations and unequal "fusibility" of various elements of granites. The granite plagioclase, in comparison with microcline, must be more stable under these conditions, and therefore it apparently lagged behind in the process of the regeneration of this granitic rock.

REVIEWS

A Textbook of Geomorphology. By PHILIP G. WORCESTER. New York: D. Van Nostrand Co., 1939. Pp. viii+565; figs. 375. $4.00.

Geomorphology: An Introduction to the Study of Landscapes. By A. K. LOBECK. New York and London: McGraw-Hill Book Co., 1939. Pp. xii+731; figs. 689. $4.50.

One of the major needs among American geological textbooks has been a modern presentation of geomorphology. Salisbury's *Physiography* appeared in 1907, Hobbs's *Earth Features and Their Meaning* was first published in 1912, while Tarr and Martin's *College Physiography* was printed in 1914. None of these volumes was strictly geomorphic, and the only critical text in English has been Cotton's *Geomorphology of New Zealand.*

This gap has now been partly filled by the appearance of two splendid volumes, carefully prepared and sumptuously illustrated. Professors Worcester and Lobeck have each made a conspicuous contribution, and teachers will find selection difficult.

The pattern of introductory courses in geology on the collegiate level seems well standardized, with semester surveys in the physical and historical aspects. Geomorphology thus appears among the advanced offerings. To the reviewer it seems unfortunate that these volumes are designed for the elementary student with little or no previous knowledge of geology. If used by those who have already had a course in general geology, there is much duplication. There is still need for a definitive volume of advanced character.

Both authors define geomorphology as that part of physiography which deals with the study of land forms and recognize that its genetic phase belongs to geology while its descriptive aspects are related to geography. Both writers regard dynamic and structural geology as separate fields, but both proceed to include extensive material on rocks, structures, and weathering. While Worcester's chapters are generally organized in terms of the land forms resulting from various agencies, Lobeck's organization is dynamic as well as topographic. The reader may well question what the scope and approach to geomorphology should be. Should the subject focus primary attention upon *surface* and consider structure and process only as necessary for explanation, or should the approach to land forms be

through the *forces* which produce them? Douglas Johnson would favor the former.[1]

Professors Lobeck and Worcester each give adequate attention to streams, ground water, glaciers, shorelines, and volcanos. Lobeck's consideration of plains, plateaus, and mountains covers 209 pages as compared with Worcester's 59. Lobeck gives almost no space to earthquakes, and Worcester has but 6 pages. Worcester devotes 32 pages to arid regions as compared with Lobeck's 2. Worcester includes material on lakes and islands, which are omitted by Lobeck. On the other hand, Worcester's treatment of coral reefs and the work of the wind is considerably briefer than Lobeck's.

Neither volume gives adequate consideration to current developments in geomorphology. Walther Penck's name does not appear in either index. Submarine canyons receive 6 pages in Worcester but only a note and a bibliography in Lobeck. Glacial stagnation is not mentioned by Lobeck and has but half a page in Worcester. Each volume considers geomorphology only as an academic subject and makes no reference to its increasing application in such fields as the Soil Conservation Service, the Bureau of Reclamation, land classification agencies such as the Tennessee Valley Authority, or the Geological Survey. Extensive bibliographies accompany each text; those in Lobeck will be especially useful to the advanced student.

Messrs. Van Nostrand and McGraw-Hill are to be congratulated on the quality of the illustrations. Worcester's volume contains 375 clear cuts, while Lobeck devotes 400 pages to illustrations. One hundred of these are full-page pictures, extending to the edge of the paper. Scores of original sketches and maps add to the understanding of the text and help to make this the most beautifully illustrated text in all geological literature.

A particular feature of Lobeck's volume is the organization. The twenty chapters are divided into 217 units, each of two or four pages, with text and explanatory illustrations on facing sheets. This often results in an arbitrary space allotment and a somewhat choppy style. Although Lobeck's contribution is 185 pages longer, Worcester has slightly more text and is also somewhat more readable.

Professors Lobeck and Worcester are each aware of the cultural values inherent in an understanding of landscapes and have described the surface

[1] Douglas Johnson, "Physiography and the Dynamic Cycle," *Science*, Vol. LXXV (1932), pp. 636–38.

of the earth in meaningful terms. The appearance of these volumes should mark a new era in the study of land forms by American students.

GEORGE B. CRESSEY

A Descriptive Petrography of the Igneous Rocks. By ALBERT JOHANNSEN. Chicago: University of Chicago Press, 1937–38. Vol. III, *Intermediate Rocks* (pp. 360; figs. 178; $4.50); Vol. IV, Part I, *The Feldspathoid Rocks;* Part II, *Peridotites and Perknites* (pp. 523; figs. 180; $4.50).

In these two volumes Professor Johannsen completes the great work heralded by the publication of Volume I in 1931. As in the earlier volumes, the rock names which he uses are those given by the original authors, but he has arranged them in orderly relation to one another within his own classification. This makes it possible for anybody, whether he is used to following Johannsen's classification or not, to use this great work as a most convenient handbook.

Both volumes are very fully illustrated, and in Volume IV there are indexes of authors, of localities, and of rock names and an appendix giving certain tables and diagrams illustrating the relationships of various dike rocks. The type, printing, and mechanics of bookmaking are excellent and of the standard that we have come to expect from the University of Chicago Press. A list of corrections for the preceding volumes is given. Most of these are matters of spelling and might originally have been typographical. The small number of corrections indicates the great care with which the copy was prepared and printed.

The scope of these volumes is pretty well indicated by the titles. The method of treatment follows consistently that in the preceding volumes. Each rock is placed in its appropriate place according to Johannsen's classification, and following the names of the rocks there are given definitions, later modifications, and current usage. In most places there are included chemical analyses and illustrations of the rock as it appears under the microscope. A special feature is the inclusion of pictures of petrographers from all over the world. To turn the pages of these volumes and to see the likenesses of professional colleagues, friends, and critics is like seeing pass in review a long column of scientific heroes.

In these, as in the preceding volumes, Johannsen has listed every igneous rock which has been carefully described in the literature. Throughout the work, the author's well-known meticulous scholarship is apparent; everything pertinent of permanent value from works of the older petrographers is included, along with very modern discussions of

differentiation and related topics. In the treatment of each major rock division Johannsen gives his own stimulating discussions, the opinions of other authors, and his own conclusions. In places he adds to the description of characteristic minerals found in certain rocks such details, for example, as the distinction between the various pyroxenes found in diorites and gabbros (Vol. III, pp. 212–13). His writings are much more than reviews of contributions to date. In masterly fashion he has analyzed the topic of "Anorthosite," to mention one example (Vol. III, pp. 201–5). In this connection, however, the reviewer regrets the absence in the Index of references to the author's own discussions, such as that on magmatic differentiation and its relation to the anorthosite referred to.

In view of the widespread increase of research in the field of ultrametamorphism, with current interest in methods of petrofabrics, and the general interest in petrogenesis, Johannsen's volumes may mark a milestone in petrology. According to general practice, Johannsen's classification is based upon the assumption that all igneous rocks are of magmatic origin and neither metasomatic, reconstructed, nor palingenetic. The origin of any rock description included by Johannsen must be attributed to the original author in every case. Johannsen has attempted to bring together in orderly fashion all those rock names which seem to have adequate credentials. He admits that some of the rocks included may not be quite orthodox petrographically, and he has deliberately left out those which he feels confident must be of metamorphic origin, even in cases where definite rock names are proposed. In this connection, of course, there have been special difficulties in relation to the monomineralic rocks, It is always questionable which of these are truly rocks and which may be considered preferably as mineral deposits. However, Johannsen has included, for the sake of completeness, some rocks which he considers of metamorphic origin. For instance, he includes, with reservations, *sagvandite* (Vol. IV, p. 459), which appears definitely to be metamorphic, judged from its mineral composition and geological relations. Perhaps less questionable is the inclusion of *eclogite* (Vol. IV, pp. 462–63). Johannsen admits it should be considered metamorphic, but nevertheless he lists and carefully describes it. Everyone will perhaps agree that it is better to include it, with a statement that it is almost certainly metamorphic, than to omit it.

With Johannsen's work in hand no other reference for the students of petrology is needed in matters of rock names, localities, authors, and essential characteristics of the rocks described. If the student were to go directly to the original non-English reference, in many cases he would be

more likely to end in error than if he were to quote Johannsen. No matter to what ends current trends in petrography may take us, the work of Johannsen will always remain a monument to the work of all petrographers to date.

T. T. QUIRKE

Physiography of the Quinnipiac-Farmington Lowland in Connecticut. By RICHARD J. LOUGEE. ("Monograph No. 7, Colby College.") Waterville, Maine, 1938. Pp. 64; pls. 15.

The study of the Quinnipiac-Farmington lowland in Connecticut was apparently undertaken as a test of two rival theories, the older of which proposes deglaciation of New England by normal, northward recession of the ice sheet, whereas the younger postulates stagnation and differential wasting.

The region studied is a long, narrow lowland within the Connecticut Valley but separated from the broader, eastern portion of the valley by a prominent trap ridge. The lowland is divisible into three sections, a northern one drained by the northward-flowing Farmington River, a central one drained by the upper portion of the southward-flowing Quinnipiac River which turns east and leaves the basin by way of a gap in a sandstone ridge, and a southern one drained by the lower Quinnipiac River on the east side of the sandstone ridge and the Mill River on the west.

Deltas and varved clays are offered as evidence of two, possibly three, lakes. The origin of the lake which is believed to have flooded the southern section of the lowland is unknown. The central lake (Southington Lake) is believed to have been formed by the till-damming of the preglacial course of the Quinnipiac River. The lake waters escaped over the crest of the sandstone ridge and cut a new channel, the deepening of which led to the draining of the lake. Farmington Lake, in the northern lowland, was impounded by the growth of alluvial fans across the lowland. As the ice front retreated to the north it exposed a gap through which the lake waters ultimately escaped.

Thus Lougee explains all of the physiographic features of the lowland—the basins, stream courses, and water gaps—as well as the distribution of the various deposits, by normal ice retreat. A large delta in the Southington basin, mapped topographically by Lougee, shows steep northern slopes, interpreted as ice-contact features, and lobate southern slopes, interpreted as indicating deposition in open water south of the ice front. The lake in which the delta was built is believed to have stretched south-

ward to a natural divide rather than to a block of stagnant ice. The assumption that stagnant ice contributed to the geomorphic development of the lowland is thus believed neither necessary nor warranted.

Uptilting of the region during ice recession is inferred largely from the southward slope of the profile drawn through the deltas of Southington Lake, although comparative profiles of river terraces are also presented as evidence.

Certain details of the history outlined by Lougee may need additional support. The proposed history of the Farmington basin is a case in point. Farmington River flows in a trench within the floor of the northern basin. The deposits west of the trench differ in composition from those on the east. Flint explained this by assuming that the trench had been occupied by a long tongue of stagnant ice against which unlike deposits from the west and east were banked. Lougee cites delta structure in the western deposits as evidence of at least local lake conditions and believes that the lake itself was the barrier that prevented coarse sediment from spreading completely across the basin. In the reviewer's opinion the evidence still admits of an explanation combining the views of Flint and Lougee, namely, that a tongue of stagnant ice lay within the confines of Southington Lake. Lougee's interpretation of the history of the Quinnipiac-Farmington lowland seems to accord with the facts presented.

The text is reasonably clear but would have benefited by the inclusion of a simple physiographic diagram of the lowland showing the principal rivers, basins, and water gaps; the included geologic map fails to satisfy this need. Line drawings interspersed through the text would also have added to the clarity. On the whole, the report will be welcomed as a valuable contribution to the problem of the deglaciation of New England.

ARTHUR D. HOWARD

Principles of Paleobotany. By WILLIAM C. DARRAH. Leiden: Chronica Botanica Co.; New York: G. E. Stechert & Co., 1939. Pp. 239. Guilders 7 (about $4.00).

This publication is Volume III of the so-called "New Series of Plant Science Books" being printed in English by Chronica Botanica Company, Leiden, Holland. By the time this review is published the final and eighth volume, *Introduction to Pollen Analysis* by G. Erdtman, will have been issued. Both of these books should prove of real importance to historical geologists and general paleontologists. Most of the others in the series probably will not.

Principles of Paleobotany is a well-integrated summary of the present status of the field of plant paleontology. The current trends in research are evaluated and discussed in their relationship to biological principles. Darrah considers the six great problems of the science to be (1) the nature of the early Paleozoic and pre-Paleozoic carbonaceous sediments, (2) the origin and ancestry of the Devonian floras, (3) the nature of the leaf, (4) the origin of seed plants, (5) the early evolution of the angiosperms, and (6) the origin of, and changes in, existing floras. All these topics are ably reviewed in chapters iv–xxi, which terminate in valuable bibliographies sufficiently up to date to include some of the writer's papers dated 1938. Chapters i–iii are excellent summaries of "Biological and Geological Principles," the "History and Aims of Paleobotany," and "Paleobotanical Techniques." Chapter xxii is a discussion of "Fossil Plants and Evolution"; and chapter xxiii, "A Reasonable Phylogeny," lists the twelve major forward steps in the phylogenetic history of the plants in terms of gross morphology. The author and subject indexes are adequate.

The book is well written. Equally important and unlike most English works printed on the Continent, it has a pleasing format and its fine typography makes it easy to read. There are, however, only seven "plates." Thus there is a lack of illustrative material, which will be an especial handicap to the nonspecialists for whose needs the book was avowedly written. This omission was probably intentional, since the *Textbook of Paleobotany* by the same author now in press (D. Appleton–Century Co.) doubtless is completely illustrated. But, if so, one may well wonder whether the present work will not lose a considerable part of its value with the publication of the text.

Some critical readers probably will get the impression that the short, two-page synoptic history of paleobotany, which has been inserted just before the Author Index, has not been prepared in an entirely disinterested fashion. It, therefore, might well have been omitted, since it serves no very useful purpose. The price ($4.00) of the work also seems rather exorbitant, especially since it is paper bound, and the author's soon-to-be-issued, more elaborate text presumably will be published in the same general price range.

CAREY CRONEIS

Geology: Principles and Processes. By WM. H. EMMONS, G. A. THIEL, C. R. STAUFFER, and IRA S. ALLISON. 2d ed. New York and London: McGraw-Hill Book Co., 1939. Pp. xii+451; figs. 468. $3.75.

The original or first edition of this book was an excellent text, but this new and thoroughly revised edition includes many improvements. It well

embodies the mature experience of the members of a department of geology who have co-operated effectively in the presentation of an elementary course in physical geology.

While the original text was noteworthy for its fine illustrations, which were largely new and unhackneyed, significant additions have been made in the way of new plates from aerial photographs and ground photographs and new block diagrams, which admirably supplement the written text.

The authors have wisely supplanted the final chapter of the first edition, which presented a brief view of historical geology, with new and pertinent material on physical geology. The introduction of new practical applications of geology, such as to soil erosion, add greatly to the general attractiveness of the volume. Actual changes in the written text as well as rearrangement of the original material make the style increasingly clear and pleasing. The chapters on diastrophism and structure have been especially improved.

This is a book which the student will find readable and the teacher will find teachable.

L. M. GOULD

Palaeozoic Formations in the Light of the Pulsation Theory, Vol. IV: *Ordovician Pulsation;* Part I: "Ordovician Formations of the Caledonian Geosyncline, with a Review and Summary of the Skiddavian Pulsation System." By AMADEUS W. GRABAU. Peking: Henri Vetch, 1938. Pp., 942; figs. 67.

Volume IV of this series considers the Ordovician formations of the Caledonian geosyncline from the viewpoint of the pulsation theory. Under this theory the phenomena of transgression and regression are explained in terms of periodic rise and fall of the sea-level.

Observations point to a simultaneous transgression of the sea in all the geosynclines, followed by a simultaneous retreat; and this, in turn, is followed by volcanic activities, orogenic movements, erosion, and accumulation of continental deposits. The numerous recurrences of such phenomena throughout the Paleozoic era are held to be rhythmic in nature; hence the term "pulsation." Recognition of such pulsations supplies a new basis for the subdivision of geologic time into periods. On this basis there would be fourteen Paleozoic periods instead of the seven recognized at present.

Coincidental with the pulsation theory is suggested the polar-control theory. It is held that the major deformations of the sial crust were due to an extra-telluric cause which was expressed in the withdrawal of a part of the crust to form a northern polar land-hemisphere to which is given the

name "Pangaea." The formation of Pangaea was accompanied by intense folding and metamorphism of the crust and was followed by continued sinking of the geosynclines and rising of the old land. These phenomena are explained as having been caused by the poleward pull of an extra-telluric (stellar) body in the position occupied by Polaris today. Thus there would have been a land-hemisphere on one side of the earth and a water-hemisphere on the other. Such a distribution of land and water could not have been maintained by an earth rotating on its axis, and eventually, probably in Jurassic time, the land mass is thought to have pulled apart, the Americas drifting westward and Eurasia drifting eastward.

Several problems of faunal evolution and distribution are explained upon the basis of the pulsation theory. The various geosynclines opened into marginal seas. In these were the centers of evolution from whence the organisms migrated into the geosynclines with the transgressing waters. When the seas subsequently retreated, the narrowing habitats led to extinction of many forms, with the survival only of "selected" faunas. These expanded into new faunal assemblages during the transgressions of the succeeding pulsation period. The respective faunas are thus linked with distinct centers of origin, and co-mingling is possible only near the confluence of geosynclines. Universal distribution is possible only for pelagic types, such as the graptolites.

In addition to these theoretical studies, the volume gives detailed consideration to Ordovician stratigraphy and the faunas of the respective horizons. Numerous columnar sections, tables, and faunal lists add to the usefulness of this extensive treatise.

RAYMOND E. JANSSEN

Les Ammonites jurassiques et crétacées. By FREDERIC ROMAN. Paris: Masson & Cie, 1938. Pp. 554; pls. 53; figs. 496. Fr. 420.

This monographic catalogue of the Jurassic and Cretaceous ammonites is one which Professor Roman has obviously attempted to make definitive for the subject. How well he has succeeded will not be determinable for perhaps a score of years, but certainly in the interim the work will prove an indispensable handbook for all students of the ammonites and for all Mesozoic stratigraphers.

The author has discussed altogether 844 genera and subgenera belonging to thirty-one families. Interestingly enough he lists only seventeen subfamilies as valid, whereas fifty-seven are placed in the synonomy.

Moreover, only thirty-four generic names are considered nonvalid, and one gains the impression that the writer was not particularly interested in taxonomic problems per se. Almost all of the genera are illustrated on the fifty-three plates. On these the figure numbers are listed consecutively from 1 to 496, most figures involving more than one picture either of the type or of a representative individual. The plate explanations conveniently face the illustrations and, in most instances, include a sutural diagnosis of the genus. Other valuable features of the publication are the phylogenetic tables for every family, which also indicate the geologic horizon marked by each genus.

The book contains an adequate Index for genera, but not for subjects; nor is there a Table of Contents. In other words, although the book is a mine of information, it will yield specific information only after careful prospecting. Fortunately it is written in lucid French, so that even a person not very familiar with the language will be able to use it. The Bibliography, which is a particularly valuable feature, includes 946 entries and is international in scope. Some rather typically North American genera, however, such as *Baculites*, receive scant space, and it is obvious that the writer has been somewhat, but probably not unduly, influenced by his geographic position.

CAREY CRONEIS

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